Bulletin

of

The Geological Society of America

VOLUME 64 May 1953 NUMBER 5

CONTENTS

	Pages
Relation of gravity anomalies to the geology of Central Vermont and New Hampshire. By Robert J. Bean	509–538
Cenozoic fill and evaporite solution in the Delaware Basin, Texas and New Mexico. By V. C. Maley and Roy M. Huffington	539–546
Mudflow of 1941 at Wrightwood, southern California. By Robert P. Sharp and Laurence H. Nobles	547–560
Systematic drainage changes in the Piedmont of North Carolina and Virginia. By William A. White	561–580
Pleistocene geology of part of central South Dakota. By Dwight R. Crandell	581–598
Short Notes	
Causes of basaltic explosions. By Harold T. Stearns	599–600
Uniformity of the earth's mantle. By Francis Birch	601–602

Subscription $15.00 per year.

Publication Office: Mt. Royal & Guilford Aves., Baltimore 2, Md.

Communications for publications should be addressed to The Geological Society of America, Dr. H. R. Aldrich, Secretary, 419 West 117 Street, New York, N. Y.

NOTICE—Four weeks notice of change of address is necessary. In accordance with the rules established by the Council, claims for non-receipt of the preceding number of the Bulletin must be sent the Secretary of the Society within three months of the date of the receipt of this number in order to be filled gratis.

Entered as second-class matter at the Post-Office at Baltimore, Md., under the Act of Congress of July 16, 1894.

Accepted for mailing at special rate of postage provided for in Section 1103 Act of October 3, 1917, authorized on July 8, 1918.

PAPERS IN PRESS FOR FORTHCOMING ISSUES

RADIATION DAMAGE IN ZIRCONS: A POSSIBLE AGE METHOD. By Patrick M. Hurley and H. W. Fairba

VERTEBRATE PALEONTOLOGY AND CONTINENTAL STRATIGRAPHY IN COLOMBIA. By R. A. Stirton

PETROLOGY OF GRANOPHYRE IN DIABASE NEAR DILLSBURG, PENNSYLVANIA. By Preston E. Hotz

GROUND-WATER BEHAVIOR IN THE HERSHEY VALLEY, PENNSYLVANIA. By R. M. Foose

TIME-STRATIGRAPHY AND THE MIOCENE EPOCH. By Martin F. Glaessner

CANONS OF LANDSCAPE EVOLUTION. By Lester C. King

WILCOX DEPOSITS IN EXPLOSION CRATERS, STEWART COUNTY, TENNESSEE, AND THEIR RELATION TO ORIGIN AND AGE OF WELLS CREEK BASIN STRUCTURE. By C. W. Wilson, Jr.

STRUCTURE OF GRAYWACKE AREAS AND TACONIC RANGE EAST OF TROY, NEW YORK. By Robert Balk

FISSILITY OF MUDROCKS. By Roy L. Ingram

NOTE—To gauge interest in the proposed Abstract Journal, to plan editions, and to estimate costs and more precise subscription price, we urge all who would be willing to pay not more than $2.00 for an annual subscription (four issues) to send a postcard indicating interest and your mailing address to

AMERICAN GEOLOGICAL INSTITUTE
2101 Constitution Ave.
Washington 25, D. C.

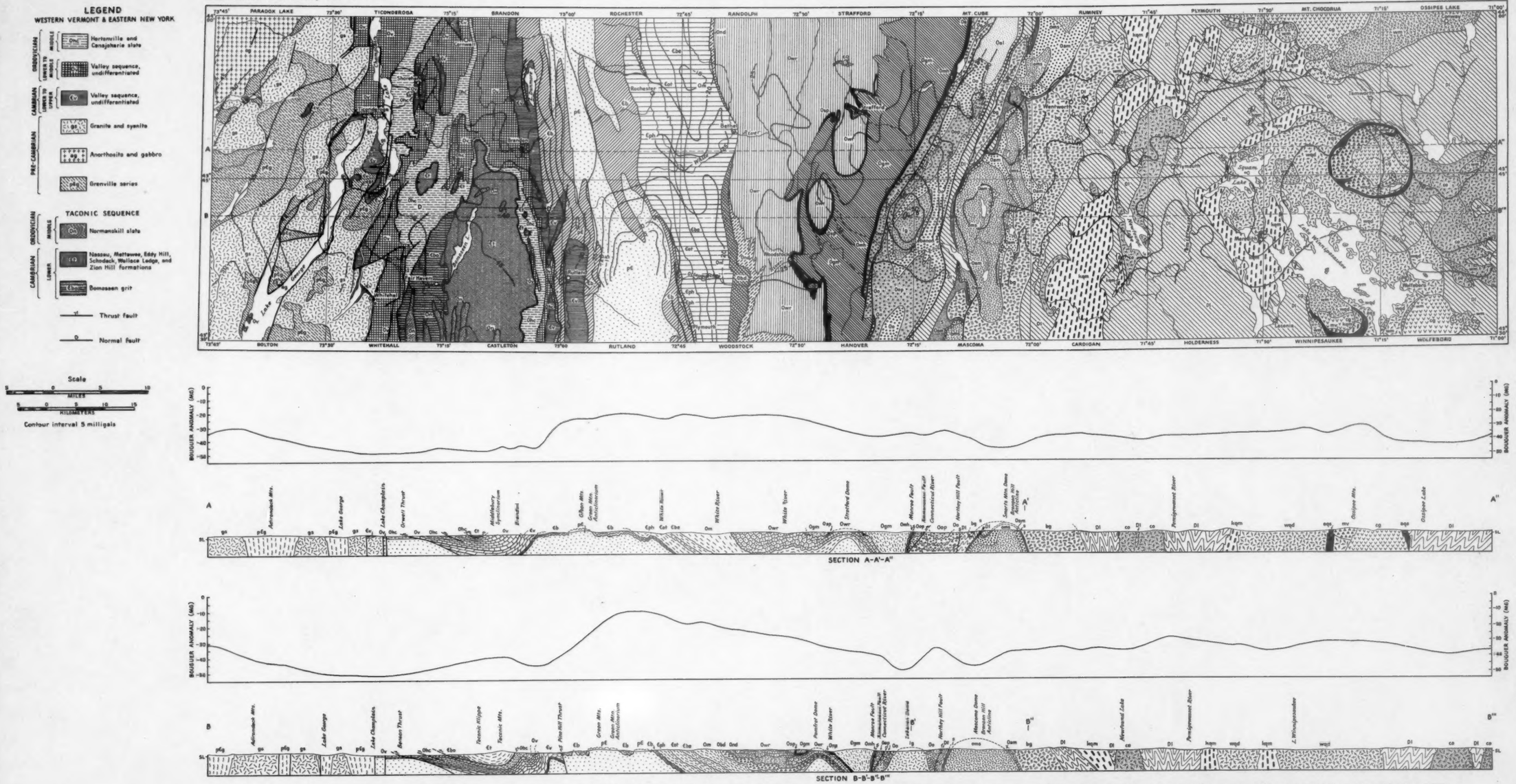
SEND NO MONEY

irba

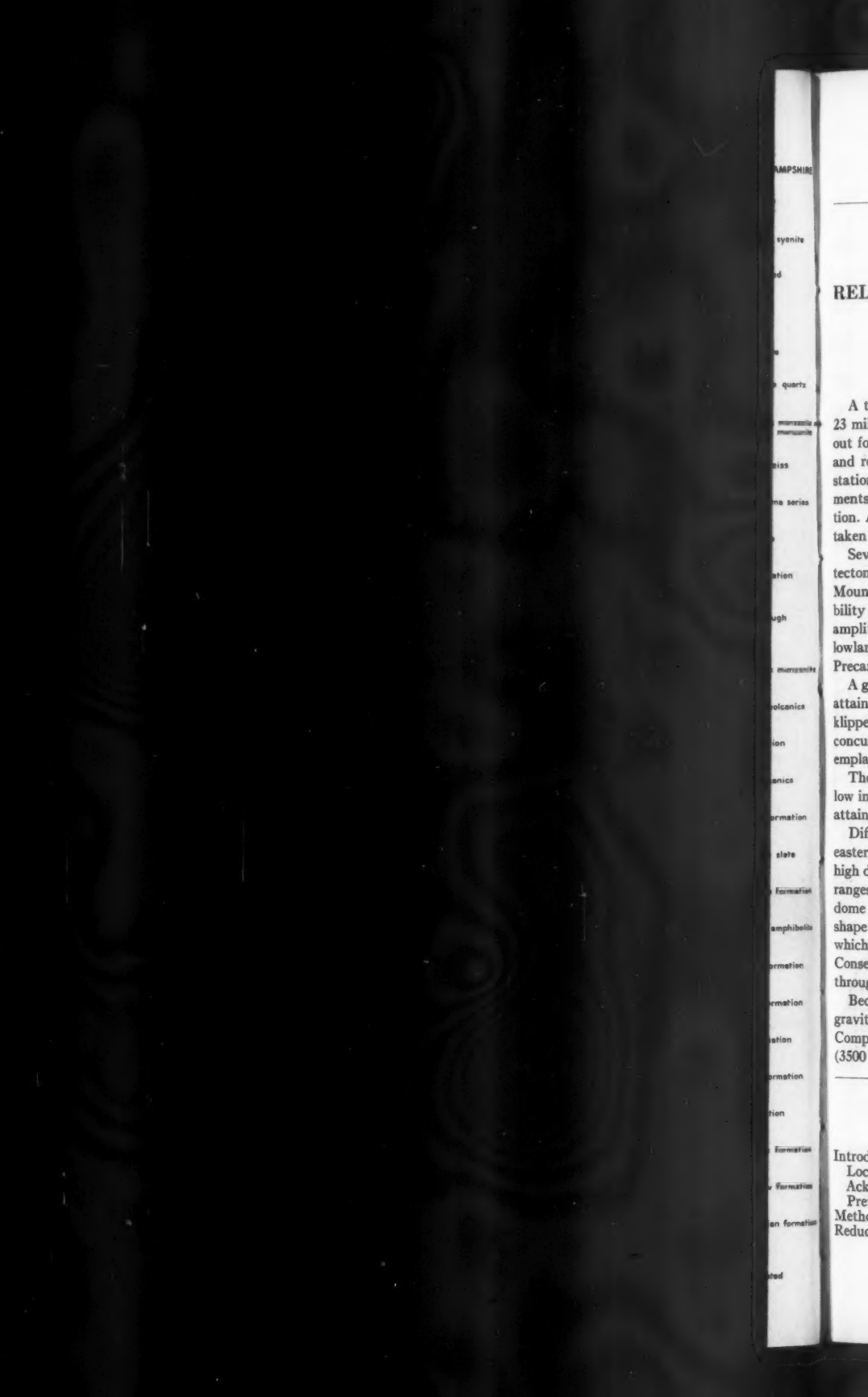
ION

Balk

sts and
2.00 for
ing ad



SIMPLE BOUGUER ISANOMALY LINES SUPERIMPOSED ON GEOLOGIC MAP OF EASTERN NEW YORK, CENTRAL VERMONT, AND CENTRAL NEW HAMPSHIRE, AND GEOLOGIC STRUCTURE SECTIONS WITH SIMPLE BOUGUER ANOMALY PROFILES



BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 64, PP. 509-530, 18 FIGS., 3 PLS. MAY 1953

RELATION OF GRAVITY ANOMALIES TO THE GEOLOGY OF CENTRAL VERMONT AND NEW HAMPSHIRE

BY ROBERT J. BEAN

ABSTRACT

A total of 1,111 gravity stations were occupied in a rectangular area approximately 137 miles long and 23 miles wide in eastern New York, central Vermont, and central New Hampshire. The survey was carried out for the primary purpose of supplementing geologic data and contributing toward the solution of local and regional structural problems. Where possible, an areal coverage method of surveying was used, and stations were located at intervals ranging from 1 to 1½ miles. The relative accuracy of the gravity measurements is approximately 1 milligal. Simple Bouguer anomalies were obtained by the usual method of reduction. Although the stations were not corrected for terrain effects for the regional analysis, this factor was taken into consideration for local analyses.

Several hypotheses are examined in an attempt to relate a large negative Bouguer anomaly field with the tectonic history of the Middlebury synclinorium, and an associated relative gravity high with the Green Mountain anticlinorium. The theory of warping of the crustal layers is applied to this region. The possibility of high-angle thrust faulting of the crust is also considered. Calculations show that the maximum amplitude of the downwarp at the base of the granitic and intermediate layers beneath the Champlain lowland is approximately 16,000 feet. This amplitude is of the same order of magnitude as the depth to the Precambrian basement in the Middlebury synclinorium.

A gravity high over the Taconic allochthone indicates that the high density slates of the Taconic sequence attain their maximum thickness in the eastern part of the thrust sheet. Because the structure of the Taconic klippe is similar to that of the Middlebury synclinorium, the rocks of the thrust sheet were probably folded concurrently with those of the synclinorium. Consequently, the Taconic thrust appears to have been emplaced prior to the close of the orogeny in the Green Mountain region.

The north-south trending regional anomalies in the Green Mountains decrease eastward to a broad gravity low in the Connecticut Valley. In the latter area, the anomaly trends swing around to the northeast before attaining east-west trends in central and eastern New Hampshire.

Differences in residual anomalies as high as −5 milligals are associated with a series of domal uplifts in eastern Vermont. These anomalies must result from a mass of low density rock which occurs beneath the high density metamorphic rocks at the surface. Calculations show that the depth to the core rocks probably ranges from 1000 to 2500 feet. An analysis of gravity minima associated with the granite core of the Lebanon dome (−13 milligals) and the core rocks of the Mascoma dome (−10 milligals) yields information as to the shape of the plutons at depth. Calculations also show that the core rocks of these domes persist to depths which are the same order of magnitude as those of the domes of eastern Vermont (12,000 to 16,000 feet). Consequently, the core rocks may be either intrusions or uplifts of a low density rock which occurs at depth throughout the region.

Because local gravity anomalies in central and eastern New Hampshire are obscured by a regional gravity low centered over the White Mountains, a residual anomaly map was prepared for this region. Computations indicate that the schists of the Littleton formation are cut off at relatively shallow depths (3500 to 8700 feet) by one or more of the magma series which occur in central New Hampshire.

CONTENTS

TEXT

	Page
Introduction	510
Location and extent of area	510
Acknowledgments	510
Previous gravity surveys	511
Method of survey	512
Reduction of data	512
Observed gravity	512
Bouguer anomaly	513
General discussion	513
Density of the lithologic units	513
Terrain corrections	513
Isostatic anomaly	515
Residual anomaly	515

	Page
Regional geology	515
Location	515
Physical features	515
Stratigraphy	516
Structure	518
Metamorphism	519
Analysis of simple Bouguer anomalies	519
Bouguer anomalies of the Green Mountain-Champlain Valley region	519
Calculation of the gravitational attraction of bodies of finite extent	525
Gravity high of Taconic Mountains	526
Eastern Vermont and western New Hampshire	526
Central and eastern New Hampshire	527
Analysis of complete Bouguer anomalies	528
Residual anomalies of eastern Vermont	528
General discussion	528
Strafford dome	529
Pomfret dome	530
Woodstock low	532
Complete Bouguer anomalies of western New Hampshire	533
General discussion	533
Lebanon dome	533
Mascoma dome	534
Stock of Conway granite	536
Conclusions	536
References cited	536

ILLUSTRATIONS

Figure	Page
1. Geographic location of area of gravity survey	511
2. Gravity stations occupied in eastern New York, central Vermont, and central New Hampshire	512
3. Regional variation of Bouguer anomaly in eastern Vermont	516
4. Structural configurations of the crust which would produce negative gravity anomaly strips	520
5. Calculated anomaly curve resulting from crustal warps in Green Mountain-Champlain Valley area	522
6. Comparison of geologic section with hypothetical configuration of Precambrian basement in Green Mountain-Champlain Valley region	524
7. Calculation of the gravitational attraction of bodies of finite extent	525
8. Calculated configuration of taconic klippe	526
9. Calculated anomalous mass associated with the Strafford dome (A-A′)	530
10. Calculated anomalous mass associated with the Strafford dome (B-B′)	530
11. Calculated anomalous mass parallel to the axis of the Pomfret low	531
12. Calculated anomalous mass associated with the Pomfret dome	532
13. Calculated anomalous mass associated with the Woodstock low	532
14. Calculated configuration of Lebanon granite along Section A-A′ across the Lebanon dome	533
15. Calculated configuration of Lebanon granite along Section B-B′ across the Lebanon dome	534
16. Calculated configuration of Lebanon granite along Section C-C′ across the Lebanon dome	534
17. Calculated configuration of Mascoma group along Section D-D′	535
18. Calculated configuration of Conway granite stock	536

Plate	Facing page
1. Simple Bouguer isanomaly lines superimposed on geologic map of eastern New York, central Vermont, and central New Hampshire	509
2. Complete Bouguer isanomaly lines superimposed on geologic map of east-central Vermont and west-central New Hampshire	532
3. Residual anomaly maps	536

INTRODUCTION

Location and Extent of Area

A gravity survey was made of a rectangular area approximately 137 miles long and 23 miles wide in eastern New York, central Vermont, and central New Hampshire (Fig. 1), between 71°00′ and 73°30′ west longitude and 43°35′ and 43°55′ north latitude. A total of 1111 gravity stations were occupied in this area during 19 weeks of the summers of 1949 and 1950 (Fig. 2).

Acknowledgments

The gravimeter was loaned to the Division of Geological Sciences of Harvard University by the Humble Oil and Refining Company. Information and assistance were kindly given by Mr. William Reid, Jr., of the exploration department of the company. The survey was carried out under the direction of Professor Francis Birch. Professor M. P. Billings and Dr. J. B. Thompson, Jr., guided the geological studies of the area. For field assistance thanks are due to Eloise E. Bean and William Diment. Acknowledgment is made to the Shaler Memorial Fund, the Cabot Fund, and the Committee on Experimental Geology and Geophysics of Harvard University for financial assistance.

The following geologists allowed the writer to use unpublished geological maps and data: Phillip Osberg, John B. Lyons, James B. Thompson, Jr., and William F. Brace.

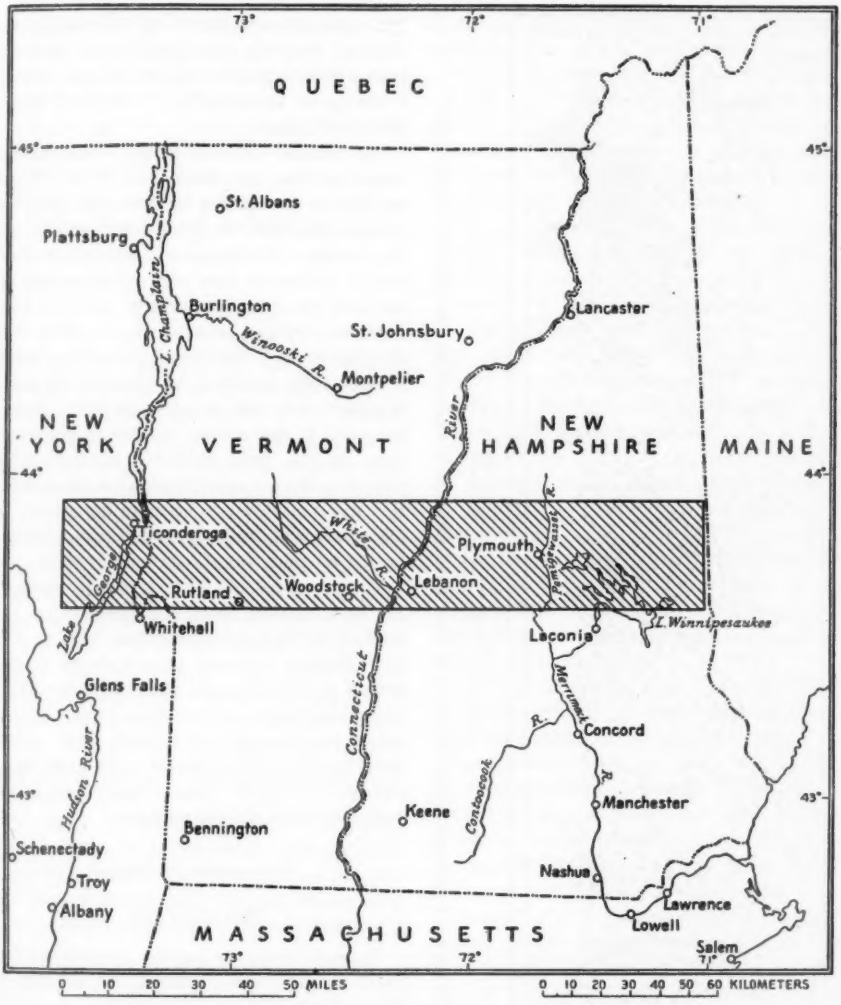


FIGURE 1.—GEOGRAPHIC LOCATION OF AREA OF GRAVITY SURVEY

The writer is indebted to Mr. F. Goldstone and Dr. W. Hafner of the Shell Oil Company for their discussion and criticism of the manuscript.

Previous Gravity Surveys

Woollard (1948) reviewed all previous gravity surveys in the New England area. Surveys which established stations in the area discussed by the writer were carried out by Woollard in 1943, 1944, and 1946. In addition, several stations were established in 1947 under the direction of Woollard during a regional survey around Lake Winnipesaukee, New Hampshire. Because all of these surveys were of a reconnaissance nature, the intervals between stations were 6 miles or more. During the summer of 1950, William Diment carried out a series of detailed traverses in the Woodsville-Barre area, Vermont.

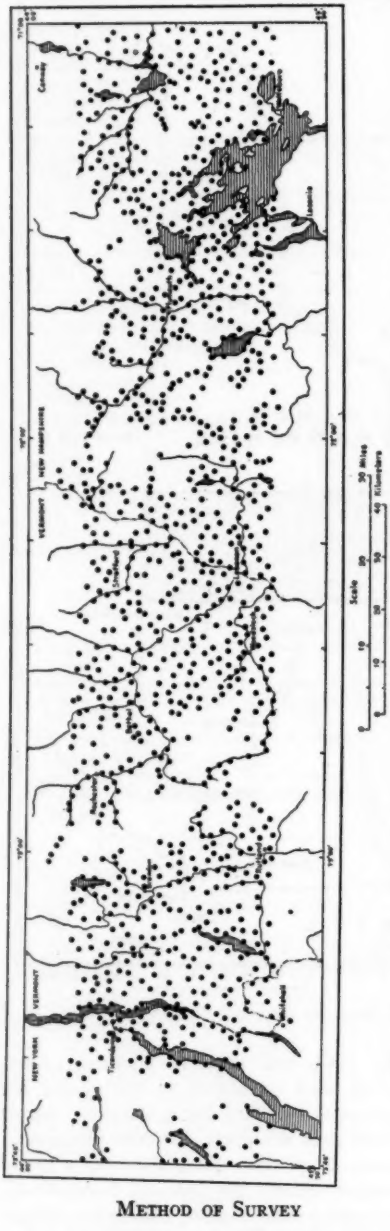


FIGURE 2.—GRAVITY STATIONS OCCUPIED IN EASTERN NEW YORK, CENTRAL VERMONT, AND CENTRAL NEW HAMPSHIRE

METHOD OF SURVEY

Gravity data were obtained by use of a Humble Oil and Refining Company gravimeter, Type I, which was mounted in a panel truck. The sensitivity constant of the instrument was obtained from the scale displacement resulting from a known change of gravity associated with a change of elevation in the Humble Bldg., Houston, Texas.

The number of stations that could be occupied in one day varied from 10 to 17, depending on the quality of the maps used, the topographic relief, the quality of the roads, and the weather. An attempt was made to cover an area of 5.5 by 6.0 miles with approximately 16 stations. Stations were usually made at road junctions, bridges, schools, etc., in order that the location and elevation of the stations could be accurately obtained. An attempt was made to collect fresh rock samples near every station occupied. It was seldom possible to place stations in the most favorable positions with respect to terrain corrections, as accurate location was considered to be more important. In a few instances, traffic interfered with readings, but at no time was it a serious problem.

To correct for the drift of the instrument, 58 base stations were established by the method of "looping" (Nettleton, 1940, p. 38). The distance between base stations ranged from 4 to 10 miles, and the time interval between occupation and reoccupation of a station varied from one-half to 4 hours. Drift curves showed that a drift of from 0.1 to 1.0 mg./hour occurred during this time. Closure corrections were applied to all base stations.

REDUCTION OF DATA

Observed Gravity

Observed values of gravity were determined from the instrument readings as follows: A barometric correction was added to the scale reading at a station, and a drift correction was applied. Any effects of temperature variations are included in the drift curves. The drift curves show that the maximum error that may be attributed to this effect is approximately 0.8 mg. Although distributed closure corrections for a single base station range from zero to 0.7 mg., only 18 per cent of the stations were given closure corrections greater than 0.2 mg. As errors arising from inaccuracies of calibration are probably small in comparison to drift and closure errors, it is believed that the ob-

served gravity is accurate to approximately 1.0 mg.

The absolute value of gravity at each station was determined by reference to the value of gravity at the U. S. Coast and Geodetic Survey gravity station No. 130, Whitehall, New York, which is 980.432 gals. Of course, the absolute value of the observed gravity is no more accurate than that of the U. S. Coast and Geodetic Survey station.

Bouguer Anomaly

General discussion.—The simple Bouguer anomaly was computed in the usual manner (Nettleton, 1940). Elevations of the stations were obtained from U. S. Geological Survey topographic maps. All of the maps except those surveyed by reconnaissance methods in eastern New York listed the elevations of most of the road junctions. Whenever possible, stations were occupied at these points; consequently, the elevations of most of the stations are probably accurate to within 10 feet. For a density of 2.67, the error in the Bouguer anomaly resulting from errors in elevation is 0.6 mg per 10 feet. Therefore, since the theoretical gravity is probably accurate to 0.1 mg. and terrain effects are not considered, the simple Bouguer anomaly at most of the stations is probably accurate to 1 mg.

In order to make the Bouguer anomalies comparable with those of other regional gravity surveys, a density of 2.67 was used in the computation of the anomaly. The regional anomaly map is constructed with a contour interval of 5 mg. (Pl. 1).

Density of the lithologic units.—In order to arrive at a mean density for each of the major lithologic units in the region surveyed, 645 rock samples were chosen for density determinations (Table 1). Although the mean densities of the units range from 2.62 (Conway granite) to 3.02 (Standing Pond amphibolite), the average density of the entire group of samples is 2.76. This value is essentially a weighted mean density with respect to the outcrop area of the lithologic units, because the number of samples of any lithologic unit is approximately proportional to the areal extent of that unit in the region surveyed.

In measuring densities, fresh, unweathered samples were chosen, and no allowance was made for porosity. A comparison between the density of dry and water-saturated samples showed a maximum difference of only 0.02, showing that the porosity or permeability of the metamorphic and igneous rocks in the region is very small, as is to be expected.

Terrain corrections.—The gravitational effect of topography was computed by means of tables prepared by Hammer (1939) which are adequate for correction of stations to 0.1 mg. A template was constructed so that terrain corrections were carried out to a radius of 48,365 feet. The addition of the terrain correction to the simple Bouguer anomaly gives the complete Bouguer anomaly.

Because the calculation of the terrain effect at a single station required from 20 to 40 minutes, only a limited number of stations were corrected. All of the base stations were corrected, in addition to one or more stations from each quadrangle which were chosen as the stations most likely to have the largest terrain corrections. The average density of 2.76 was used in the computations.

Terrain corrections were also made for all stations in certain local areas where the gravity data were subjected to detailed analysis. The density of the country rock in the area was used in these calculations. Because of the time involved in making the corrections, a method given by Bickel (1948) was used to obtain corrections for the outer four zones (radius greater than 8,578 feet).

A total of 379 stations were corrected for terrain effects. The corrections for the base stations together with the corrections carried out for detailed analyses should indicate the general magnitude of the terrain corrections for all the stations, whereas the stations chosen for their large terrain effects should yield maximum values.

The corrections were distributed as follows (except maximum corrections):

Number of stations	*Correction (mg.)*
2	4.0–4.9
3	3.0–3.9
48	2.0–2.9
169	1.0–1.9
135	0.0–0.9

TABLE 1.—DENSITIES OF LITHOLOGIC UNITS

Lithologic Unit	Number of Samples	Mean Density	Range of Density
White Mountain magma series			
Conway granite	4	2.61	2.60–2.62
Conway granite (Billings, 1928)	3	2.62	2.60–2.65
Moat volcanics (Mafic)	8	2.90	2.85–2.93
New Hampshire magma series			
Concord granite	21	2.67	2.61–2.75
Winnipesaukee quartz diorite	49	2.73	2.67–2.83
Kinsman quartz monzonite	11	2.70	2.68–2.74
Bethlehem gneiss	18	2.75	2.71–2.77
Oliverian magma series			
Mascoma group			
Granite & quartz monzonite	13	2.68	2.65–2.69
Granodiorite	6	2.71	2.69–2.72
Lebanon group			
Lebanon granite	5	2.66	2.64–2.69
Lebanon border gneiss	6	2.79	2.72–2.89
Littleton formation			
Medium-grade metamorphism	8	2.76	2.71–2.79
High-grade metamorphism	76	2.82	2.70–2.96
Clough formation	3	2.68	2.67–2.70
Ammonoosuc volcanics			
amphibolite	7	2.96	2.92–3.08
felsic	8	2.73	2.72–2.75
Albee Formation	3	2.73	2.71–2.75
Post Pond volcanics	26	2.85	2.69–3.12
Orfordville formation	14	2.75	2.69–2.83
Hardy Hill quartzite member	3	2.68	2.65–2.69
Meetinghouse slate	5	2.75	2.72–2.80
Gile Mountain formation	39	2.76	2.69–2.91
Standing Pond amphibolite	6	3.02	2.79–3.14
Waits River formation	34	2.76	2.68–2.92
Northfield formation	23	2.83	2.71–3.01
Barnard gneiss	7	2.70	2.64–2.75
Missisquoi formation	13	2.81	2.72–3.00
Bethel formation	8	2.84	2.76–3.03
Ottauquechee formation	11	2.79	2.73–2.89
Pinney Hollow formation	18	2.83	2.78–2.97
Basal Cambrian formations			
Tyson and Grahamville formations	8	2.85	2.82–2.89
Dolomite	2	2.86	2.83–2.88
Cheshire quartzite	4	2.64	2.57–2.73
Moosalamoo phyllite	4	2.75	2.73–2.78
Pinnacle arkose	3	2.67	2.66–2.68
Precambrian			
Rocks in			
Green Mountain anticlinorium	11	2.74	2.68–2.82
Gneiss in Chester dome (Thompson, 1950)	7	2.69	2.66–2.73
Hortonville slate	13	2.73	2.70–2.80
Canajoharie slate	3	2.70	2.65–2.74
Cambro-Ordovician dolomite (Champlain Valley)	34	2.84	2.80–2.86
Cambro-Ordovician limestone (Champlain Valley)	30	2.73	2.67–2.80
Potsdam sandstone	14	2.64	2.58–2.68

TABLE 1.—*Concluded*

Lithologic Unit	Number of Samples	Mean Density	Range of Density
Precambrian (Adirondack Mountains)			
Grenville formation			
Gneiss	25	2.84	2.70–3.06
Limestone	2	2.73	2.72–2.74
Quartzite	1	2.62	
Granite and syenite	11	2.70	2.63–2.79
Anorthosite	3	2.72	2.68–2.75
Gabbro	3	2.97	2.91–3.04
Taconic Sequence			
Slate	17	2.81	2.72–2.84
Bomoseen grit	1	2.72	
Phyllite of Nassau formation	1	2.82	

Maximum corrections were distributed as follows:

Number of stations	*Correction (mg.)*
1	9.0
1	6.0
3	5.0–5.9
2	4.0–4.9
6	3.0–3.9
6	2.0–2.9
3	1.0–1.9

Since most of the terrain corrections are less than 3.0 mg, the failure to consider this factor in the regional analysis will not appreciably affect the interpretation.

As the corrections for curvature and indirect effect varied only a few tenths of a milligal over the entire area, these effects were neglected.

Isostatic Anomaly

The Hayford isostatic anomaly with a depth of compensation of 113.7 km. may be determined for stations in the region by adding to the Bouguer anomaly an isostatic correction obtained from U. S. Coast and Geodetic Survey pendulum stations in the New England area. As the isostatic correction ranges from +23 mg. in the west to +5 mg. in the east, an isostatic anomaly map would be quite similar to the Bouguer anomaly map.

Residual Anomaly

In many cases, relatively small anomalies which reflect variations in density in the surface or near-surface rocks are obscured by large gravity trends related to regional tectonic features. To study the small anomalies, these larger trends were removed by Wilson's (1941) method (Fig. 3). In eastern Vermont, the predominant east-west regional trend was approximated with sufficient accuracy by a straight line, and the other set of points showed that no regional variation occurred perpendicular to this trend. Subtraction of the regional effect at each station yielded the residual anomaly.

REGIONAL GEOLOGY

Location

This geological summary is concerned primarily with the quadrangles in which gravity stations were occupied. The region lies between 43°30′ and 43°60′ north latitude, and 73°30′ and 71°00′ west longitude (Pl. 1).

Physical Features

The principal highlands west of the Connecticut River are the Adirondack Mountains of eastern New York, which attain elevations of 2000 feet; the Green Mountains, where peaks rise to 4000 feet; and the Taconic Mountains, which lie in the Champlain lowland west of the Green Mountain front. The highlands of western New Hampshire trend northeastward and rise to elevations of 4000 feet in the southern parts of the White Mountains of cen-

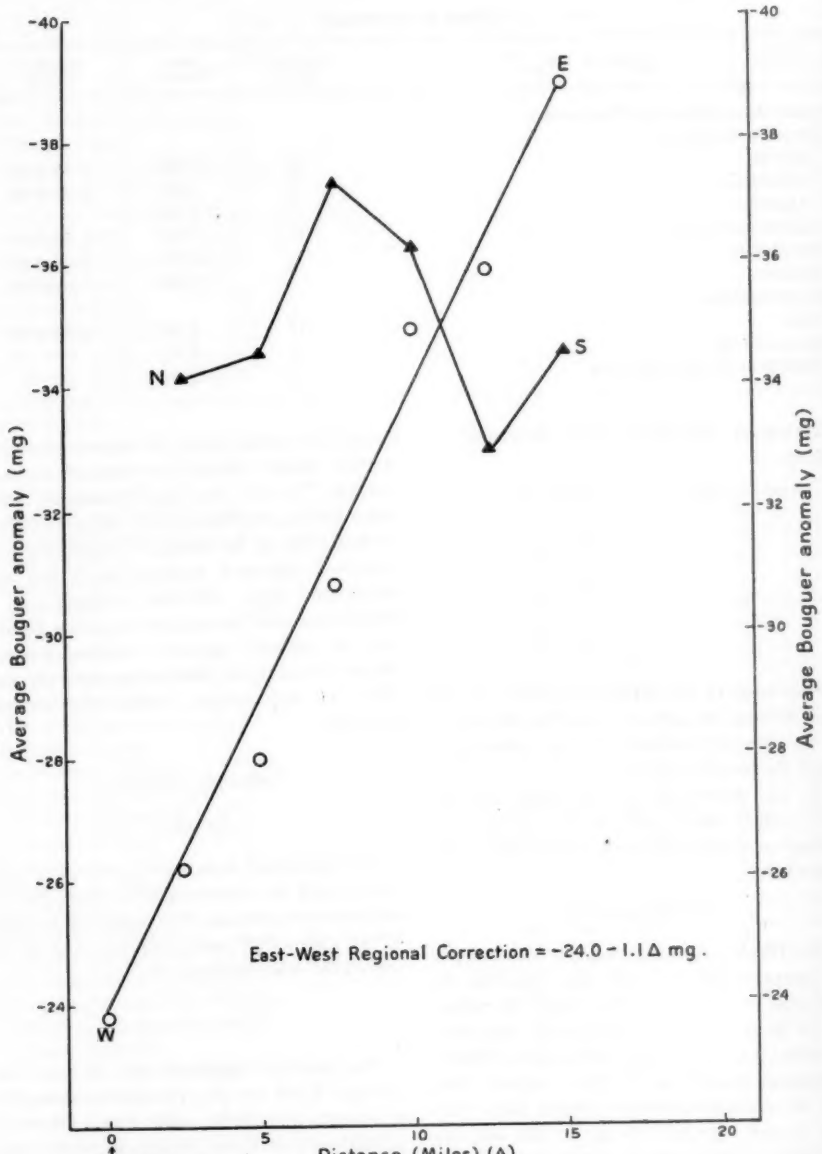


FIGURE 3.—REGIONAL VARIATION OF BOUGUER ANOMALY IN EASTERN VERMONT

tral New Hampshire. The Winnipesaukee lowland of eastern New Hampshire is occupied by many lakes and highlands of limited extent, such as the Ossipee Mountains.

Stratigraphy

Precambrian rocks occur in the Adirondack Mountains and in the core of the Green Moun-

tain anticlinorium. In the former area, the Grenville series, which is composed principally of sillimanite and pyroxene gneisses, graphitic marble, and quartzite, is intruded by anorthosite, gabbro, granite, and syenite. Rocks in the latter area consist of highly metamorphosed gneiss, schist, dolomite and quartzite.

Basal Cambrian formations in the Green Mountains are composed of arkose, quartzite, phyllite, and dolomite. The overlying Cambro-Ordovician sequence in western Vermont is markedly different from that in eastern Vermont. Predominantly calcareous rocks occur in the west in the form of siliceous dolomite and limestone with minor amounts of quartzite and slate, whereas a predominantly argillaceous facies occurs in the east. As the Pinney Hollow-Ottauquechee-Bethel-Missisquoi sequence of eastern Vermont consists principally of sericite-chlorite schist with subordinate quartzite and greenstone, Thompson (1950) suggests that these formations may be correlated with the grits and green, purple, and black slates of the Taconic sequence. Although the Bomoseen grit and the sericite-chlorite schist of the Nassau formation occur at the base of the Taconic section, their mutual relations are unknown (Kaiser, 1945).

The hornblende gneiss and amphibolite of the Barnard formation, which overlies the Missisquoi formation in eastern Vermont, may be metamorphosed volcanics. A probable Middle Ordovician age is assigned to the Northfield phyllite and the massive, gray-blue, arenaceous marble of the Waits River formation on the basis of meager fossil evidence. Therefore, these formations may be tentatively correlated with the Hortonville-Canajoharie slates of the Champlain lowland and with the Normanskill slate of the Taconic sequence.

An excellent horizon marker, the Standing Pond amphibolite, occurs near the top of the Waits River formation. The thickness of the Cambro-Ordovician section in western Vermont is estimated to be 8000 feet; the thickness in eastern Vermont to the top of the Waits River formation is conservatively estimated to be 20,000 feet (Chang, 1950). The Taconic section is approximately 3500 feet thick (Fowler, 1949).

Upper Ordovician (?) formations which occur in eastern Vermont and western New Hampshire are, in order of decreasing age, as follows.

(1) Gile Mountain formation—chiefly quartz-mica schist which locally contains garnet, staurolite, and sillimanite.
(2) Orfordville formation—predominately phyllite, slate, and quartzite which is locally metamorphosed to biotite, garnet, staurolite, and kyanite schists.
(3) Post Pond volcanics—amphibole gneiss, biotite schist, amphibolite (metamorphosed felsic and mafic tuffs).
(4) Albee formation—quartzite and slate locally metamorphosed to mica schists and quartz-mica schists.
(5) Ammonoosuc volcanics—amphibolite, biotite gneiss, hornblende gneiss (metamorphosed tuffs).

The thickness of these formations is estimated to be 11,000 feet in this area.

The Clough and Fitch formations have been dated as Lower to Middle Silurian on the basis of fossil evidence. The formations contain quartzite, quartz conglomerate, quartz-mica schist, arenaceous marble, and lime-silicate granulite. Fossil evidence is also the basis for the Lower Devonian age of the Littleton formation. The predominate mica schist and quartz-mica schist of the formation may contain porphyroblasts of biotite, staurolite, garnet or sillimanite, dependin9 on the metamorphic facies. Quartzite, lime-silicate granulite, biotite gneiss, and amphibolite are common. The formation is from 4000 to 11,000 feet thick (Chapman, 1939; Moke, 1945). The Mississippian (?) Moat volcanics are the extrusive phase of the White Mountain magma series and consist of interbedded flows, tuffs, and breccias of rhyolite, trachite, andesite, and basalt.

Plutonic rocks which occur in New Hampshire are as follows.

(1) Fairlee quartz monzonite (Upper Ordovician ?)
(2) Lebanon group (Middle Devonian ?)—pink granite surrounded by a dark, mafic border phase.
(3) Oliverian magma series (Middle Devonian ?)—composition ranges from quartz diorite to granite.
(4) New Hampshire magma series (late Devonian ?)—Bethlehem gneiss ranges from granite to granodiorite. The Kinsman

quartz monzonite, Winnipesaukee quartz diorite, and Concord granite are other members of the series.

(5) White Mountain magma series (Mississippian ?)—alkalic rocks that are predominately felsic, but may range from gabbro to granite.

Structure

Stratigraphic evidence indicates that the orogeny in the Green Mountain region is late Ordovician in age (Taconic). In central and western Vermont, the sediments were folded into the broad Green Mountain anticlinorium and the adjacent Middlebury synclinorium. A break thrust—the Pine Hill thrust—developed along the flexure common to these major structural units. The eastward dip of the thrust plane, together with the overturning of the anticlinorium toward the west, indicates that during the orogeny the upper parts of the crust moved west relative to the deeper parts. Moreover, rocks of the Middlebury synclinorium are thrust westward over the early Paleozoic rocks of the foreland, which are extensively normal faulted. Cady (1945) reports that the planes of the major low-angle thrust faults are nearly parallel to the overlying and underlying strata for long distances. The synclinorium plunges south and is covered by the Taconic allochthone in the southern part of the area.

Fowler (1949) and Kaiser (1945) have reviewed the structural and stratigraphic evidence and conclude that the Taconic sequence is structurally a gigantic klippe. The edge of the Taconic thrust has not been definitely located everywhere, because throughout most of its northern extent argillaceous deposits of the allochthone overlie lithologically similar argillaceous beds of the autochthonous series. The rocks of the Taconic sequence are isoclinally folded along north-south axes and overturned to the west. It is generally believed that the Taconic thrust sheet originated to the southeast, but no root zone has yet been discovered.

A series of domes trend north-northeast through eastern Vermont. Two of these, the Pomfret and the Strafford domes, are within the area of the survey. They are outlined by the Standing Pond amphibolite, the Waits River formation is exposed in the center of the domes, and they are wholly or partially surrounded by the Gile Mountain formation. The domes are roughly symmetrical and appear to have originated through essentially vertical movements. Reverse drag folds in the Chester dome south of the Woodstock quadrangle support this method of origin. Erosion has exposed the Precambrian core in the Chester dome, and the formations around its flanks have undergone tectonic thinning to approximately a tenth of their apparent thickness in the homoclinal sequence (Thompson, 1950). Many minor synclines and anticlines occur in the Waits River formation between the eastward dipping homoclinal sequence on the west and the domes on the east.

Three thrust faults, the Monroe, Ammonoosuc, and Northey Hill, are located between the domes of Vermont and the Bronson Hill anticline, a series of north-northeast trending en echelon domes in western New Hampshire. The Oliverian magma series in the cores of the domes is concordant with the overlying formations. Several hypotheses concerning the origin of the domes have been presented. They have been considered as syntectonic phaccoliths, horizontal sheetlike intrusions which have been subsequently domed upward (Hadley, 1942), or laccolithic intrusions (Chapman, 1939).

The relation of the Lebanon dome to the others is obscure. Granite and border gneiss appear to have been forcefully injected into the Orfordville formation, and the dome is overturned toward the south-southeast.

Synclines are present both west and east of the Bronson Hill anticline, but structural details cannot be satisfactorily determined in the Littleton formation of central and eastern New Hampshire, partly because of the intense folding and intrusion, but principally because of the lack of distinctive stratigraphic units.

The Bethlehem gneiss and Kinsman quartz monzonite form essentially concordant sill-like plutons which dip steeply to the east or are vertical. Hadley (1942) and Billings (1945) believe that the Bethlehem gneiss was intruded as a horizontal sheet and subsequently domed upward by the intrusion of the Oliverian magma series. This mechanism would explain the isolated body of gneiss which occurs west of the Bronson Hill anticline in the Mt. Cube

quadrangle. The Winnipesaukee quartz diorite and the Concord granite show both concordant and discordant relations with the surrounding rocks.

Ring-dike complexes are peculiar to the rocks of the White Mountain magma series. A prominent example in the Ossipee Mountains is outlined by an intrusion of the Albany quartz syenite, which probably stoped upward along arcuate fracture zones. This resulted in the subsidence of the central block and in the preservation of the Moat volcanics (Billings, 1945). Several stocks of the White Mountain magma series are present in the region. The White Mountain batholith, the southern portion of which lies in this area, is composed of a complex of coalescing ring dikes and stocks.

As the pre-Pennsylvanian White Mountain magma series cuts the Lower Devonian Littleton formation, the orogeny in New Hampshire is believed to have occurred during the Upper Devonian (Acadian).

Metamorphism

High-grade (sillimanite) metamorphic rocks occur in the Precambrian core of the Green Mountain anticlinorium, in the Grenville series, and in the Littleton formation east of the Bronson Hill anticline. In eastern Vermont, sillimanite bearing rocks are associated with the domal uplifts. The intensity of metamorphism decreases eastward from the domes to a low-grade (chlorite) facies that occurs west of the Ammonoosuc fault, and then increases to medium-grade (biotite, garnet, staurolite, and kyanite) facies in western New Hampshire. To the west of the Vermont domes, middle-grade rocks predominate, but these are replaced by a low-grade facies in the Middlebury synclinorium and in the Taconic klippe. The Paleozoic rocks of the foreland have not been metamorphosed.

Analysis of Simple Bouguer Anomalies

Bouguer Anomalies of the Green Mountain-Champlain Valley Region

The present gravity study of the Green Mountain-Champlain Valley region confirms the earlier reconnaissance work in New England by Woollard (1948). A broad gravity low in the Champlain Valley lowland attains a minimum value of −57 mg. near Whitehall, New York (Pl. 1). The anomalies increase gradually both east and west of the axis of the gravity low. Maximum anomalies of −30 mg. are found in the eastern Adirondack Mountains, whereas relatively high anomalies occur over the Taconic Mountains. To the east of Rutland and Brandon, Vermont, the anomalies abruptly increase to a maximum of −3 mg. along the axis of the Green Mountains. In the southern part of the area, the anomalies increase from −42 mg. to −3 mg. in a horizontal distance of 8 miles, a gradient of approximately 5 mg. per mile. The Champlain low is the northward continuation of the Hudson Valley low of Longwell (1943), and the Blue Ridge low of Woollard (1939; 1943) and Nettleton (1941).

The Green Mountain "high" is the northward continuation of a gravity high which occurs east of the Blue Ridge low in the Appalachian area.

The regional anomaly patterns do not result from any systematic variation in the densities of the lithologic units (Table 1). The Champlain Valley low embraces the Adirondack granitic intrusives (density, 2.70), the Grenville gneiss (2.84), and the limestones (2.73) and dolomites (2.84) of the Champlain Valley. On the other hand, the Green Mountain high overlies the Precambrian core of the Green Mountains (2.74), basal Cambrian formations (2.67–2.85), and the lower portions of the Cambrian section of eastern Vermont. Consequently, the mass distribution which causes the regional anomalies must be located at depth.

Several hypotheses have been presented to explain these regional anomalies. Large, linear gravity lows have been considered as resulting from

(1) a basin of low density sediments overlain by thrust sheets of greater density (Fig. 4A);
(2) the depression into the substratum of the crustal member overridden by the thrust sheets by
    (a) thrust faulting of the crust (Fig. 4C);
    (b) depression of the crust by the excess mass of thrust sheets (Fig. 4B);
(3) the downbuckling of a portion of the earth's crust (Fig. 4D);

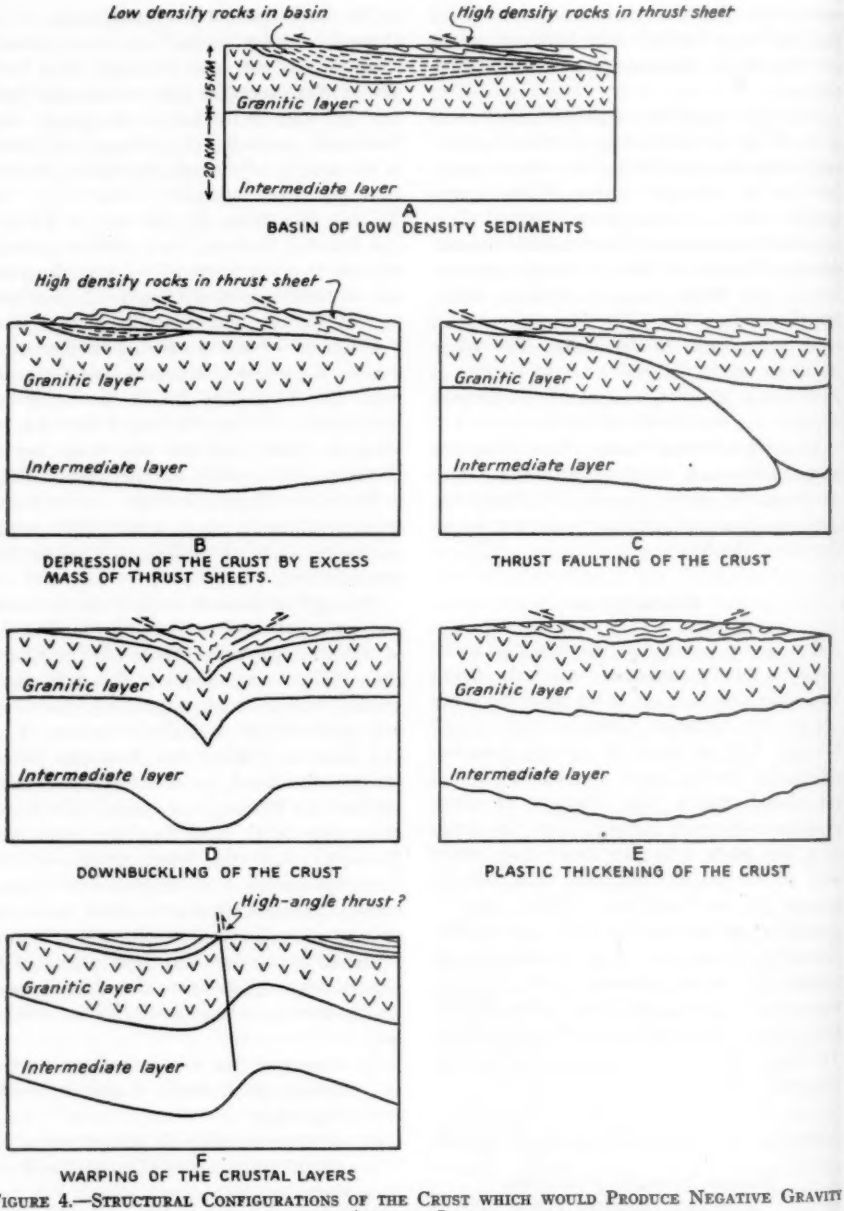


FIGURE 4.—STRUCTURAL CONFIGURATIONS OF THE CRUST WHICH WOULD PRODUCE NEGATIVE GRAVITY ANOMALY STRIPS

(4) plastic thickening of the crust as a result of horizontal compression (Fig. 4E);
(5) warping of the subcrustal layers (Fig. 4F).

The first of these theories does not apply to the Champlain Valley region, because the average density of the limestones and dolomites of the foreland is as great as the density of the

limestones and dolomites of the thrust sheets. In addition, the density of the calcareous rocks is as great as that of the schists and gneisses in the Adirondack and Green mountains.

With respect to the second theory, the axis of the Champlain low coincides approximately with the western limits of the thrust sheets. This zone of overthrusting may be the result either of faults that affect the entire crust, or of superficial faults that cut only the upper portions of the crust. The thrusts may be related to a fault system that dips eastward at a low angle near the surface but attains steeper dips at depth. If the entire crust was sheared and thrust faulted along these zones, the underthrust portion would have been pushed into subcrustal depths. Therefore, the axis of the negative anomaly band which would result from the underthrust mass would lie east of the trace of the thrust planes; consequently, this theory does not explain the existing anomaly pattern.

If the thrusts are essentially superficial features of crustal deformation, it is possible that during the orogeny a large mass of rock was added to the crust in the Champlain Valley region by large-scale overthrusting. This excess mass would probably depress the crust in the region in a manner similar to the deformation of an elastic plate under a superimposed load. However, the negative anomaly strip resulting from such a process would tend to be eliminated if the major part of the transported mass were removed by erosion. The crustal reaction in such a case would be similar to the recoil of regions of continental glaciation. As the essentially undeformed Paleozoic foreland sequence is exposed in the Champlain lowland, and the gently dipping thrust sheets are confined to the eastern parts of the region, any large overthrust mass which once might have covered the entire lowland has been largely removed by erosion. Therefore, it is believed that this theory does not explain the existing anomalies.

The third hypothesis considers the possibility that a crustal downbuckle similar to that postulated by Vening Meinesz (1934) is present beneath the negative strip. The fourth hypothesis calls for a plastic thickening of the crust to compensate for the shortening produced by mountain building. The principal difference in the two theories involves the relative competency of the crust to transmit stress, which is an unknown factor. According to these theories, the zone of most intense orogenic deformation should coincide with the negative zone. This is definitely not the case in the Champlain Valley, because the axis of the gravity low lies in a region composed of flat-lying, essentially undeformed Cambro-Ordovician sediments. Even the rocks in the thrust sheets to the east are not greatly deformed, and metamorphism is of the lowest grade.

The fifth hypothesis deals with warping of the subcrustal layers. Application of Glennie's (1933) crustal warp hypothesis to the Appalachian gravity problem has been considered by Woollard (1939) and by Longwell (1943). According to this theory, broad negative anomalies are present over areas of crustal downwarp, while positive anomalies are present over upwarped areas where the dense subcrustal material approaches closer to the surface.

Application of the theory of crustal warping to the Green Mountain-Champlain Valley anomaly strips helps to clarify some of the larger structural problems of this region. However, the theory could be modified to include the possibility of thrust faulting as proposed by Woollard (1939; 1940). The broad Green Mountain anticlinorium and Middlebury synclinorium fit into such a structural picture, and the rarity of igneous intrusions in the area tends to support the hypothesis that no deep "root zone" was formed during the deformation. In fact, because of the presence of the relative gravity high over the Green Mountains, Woollard (1944) noted that the region illustrated the case of mountains without roots.

To calculate the amount of crustal warping necessary to produce the observed anomalies, certain simplifying assumptions must be made.

(1) Seismic data show that a "granitic" layer approximately 15 km. thick and an "intermediate" layer approximately 20 km. thick underlie New England (Leet, 1941). These values may be taken to represent the average thickness of the layers beneath the entire New England area. The assumption is made that the average thickness of the crustal layers has not changed since early Paleozoic time. Since subsequent intense orogeny has occurred throughout much of the region, the assumed thickness is probably too large.

(2) The density of the granitic layer is as-

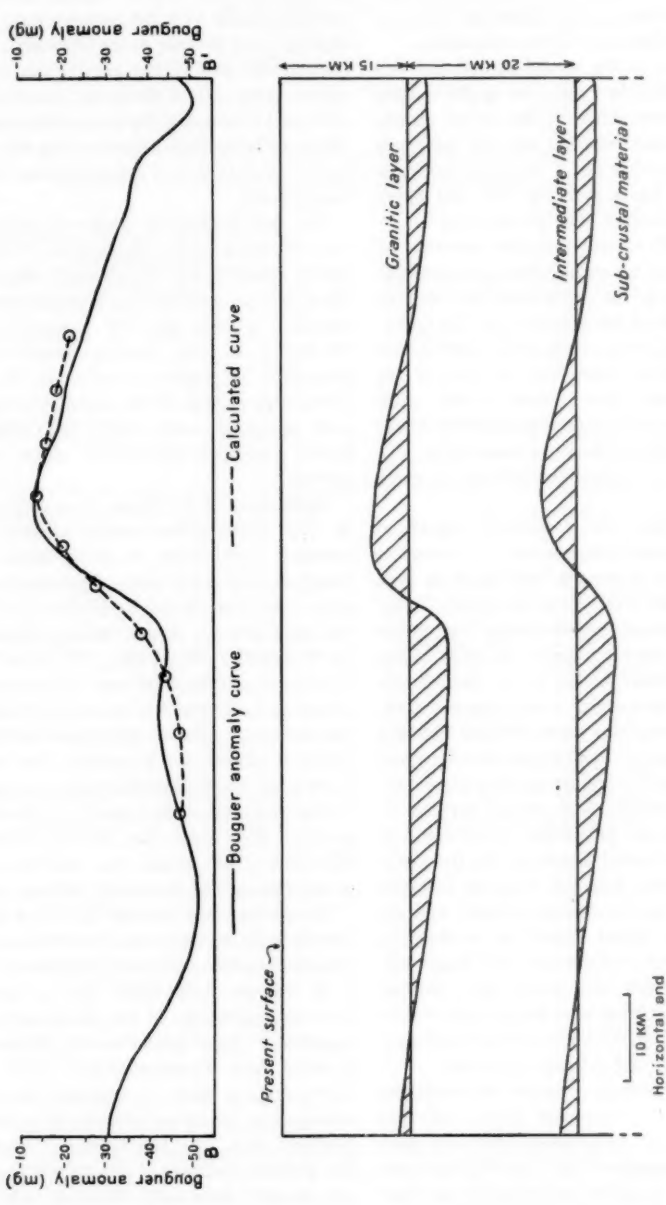


FIGURE 5.—CALCULATED ANOMALY CURVE RESULTING FROM CRUSTAL WARPS IN GREEN MOUNTAIN-CHAMPLAIN VALLEY AREA (B-B', Pl. 1)

sumed to be 2.7, that of the intermediate layer to be 3.0, and that of the subcrustal material to be 3.3. Therefore, the density differences resulting from crustal warps at the base of either the granitic or intermediate layer is 0.3.

(3) It is assumed that, during the crustal warping, both of the layers were deformed in the same manner.

(4) The mass anomalies produced by the crustal warps are assumed to extend in an infinite direction parallel to the strike of the anomalies and the geologic structure. Because the gravity anomalies continue north and south of the region surveyed, no appreciable error results from this assumption.

A method of computing the gravimetric effects of two-dimensional masses given by Hubbert (1948) was used to calculate the form and magnitude of crustal warps which would produce the observed gravity anomalies (Fig. 5). The calculated curve which best fitted the data could be computed with accuracy for only the central part of the gravity profile, since unknown effects became significant beyond this region.

It should be emphasized that, although the calculated gravity effect of an assumed mass distribution closely approximates the observed gravity curve, the inherent ambiguity of gravity interpretation precludes any assertion that the assumed mass distribution is correct. Whether the interpretation is correct or not depends upon the degree to which the various unknowns at depth are correctly estimated. Values for the unknown quantities must be supplied by geological analysis.

If we assume that the granitic layer which existed during Cambro-Ordovician time was 15 km. thick and consisted of Precambrian rocks, and that this surface was deformed during the orogeny in a manner similar to the warps within the crust (Fig. 6), the maximum amplitude of the crustal downwarp should be the same order of magnitude as the maximum depth of the Precambrian basement in the Champlain Valley. The maximum amplitude of the crustal downwarp in this area is computed to be approximately 5 km. or 16,000 feet. This value may be compared with the depth of an extrapolated east-west geologic cross section located a mile north of Brandon, Vermont (Cady, 1945), which is estimated to be 12,500 feet. However, because the Middlebury synclinorium plunges southward in the area, the depth would be greater along the lines of the gravity section. Of course, both of the estimates as to the depth of the basement are subject to large errors in view of the simplifying assumptions on which they are based, but the fact that the results are of the same order of magnitude is encouraging.

The maximum depth of the crustal downwarp in the Champlain Valley region occurs immediately west of the Green Mountain front. In addition, the general structure of the crustal warps confirms geologic evidence that the upper crustal units moved westward relative to the lower units. The break thrusts which developed along the western Green Mountain front and the thrust sheets resulted from these differential movements. Possibly the Pine Hill thrust is related to a zone of faulting which extends downward along the steeply dipping limb of the major flexure below the Green Mountain front.

The question arises whether such crustal warps and their accompanying anomalies would persist through long periods of geologic time without isostatic adjustment occurring. It seems probable that any large anomaly would be reduced until the upward (or downward) force resulting from the anomaly were balanced by the strength of the crust. It should be possible for such anomalies to exist for essentially an indefinite period of time, providing, of course, that the crustal configurations are not disturbed by a new orogenic cycle. The magnitude of the anomaly which remains would depend on the strength of the earth's crust.

There is no evidence that either the Appalachian orogeny or the Triassic disturbance appreciably affected the Green Mountain area. If the orogeny occurred at the close of the Ordovician period, the intense Acadian orogeny of New Hampshire should have greatly disturbed the region. However, there is no evidence that the Green Mountain region as a whole was affected by two periods of orogeny. Broad crustal movements during which subcrustal flow possibly takes place, such as glacial depression and recoil, and epeirogenic movements, would probably not affect the crustal configurations.

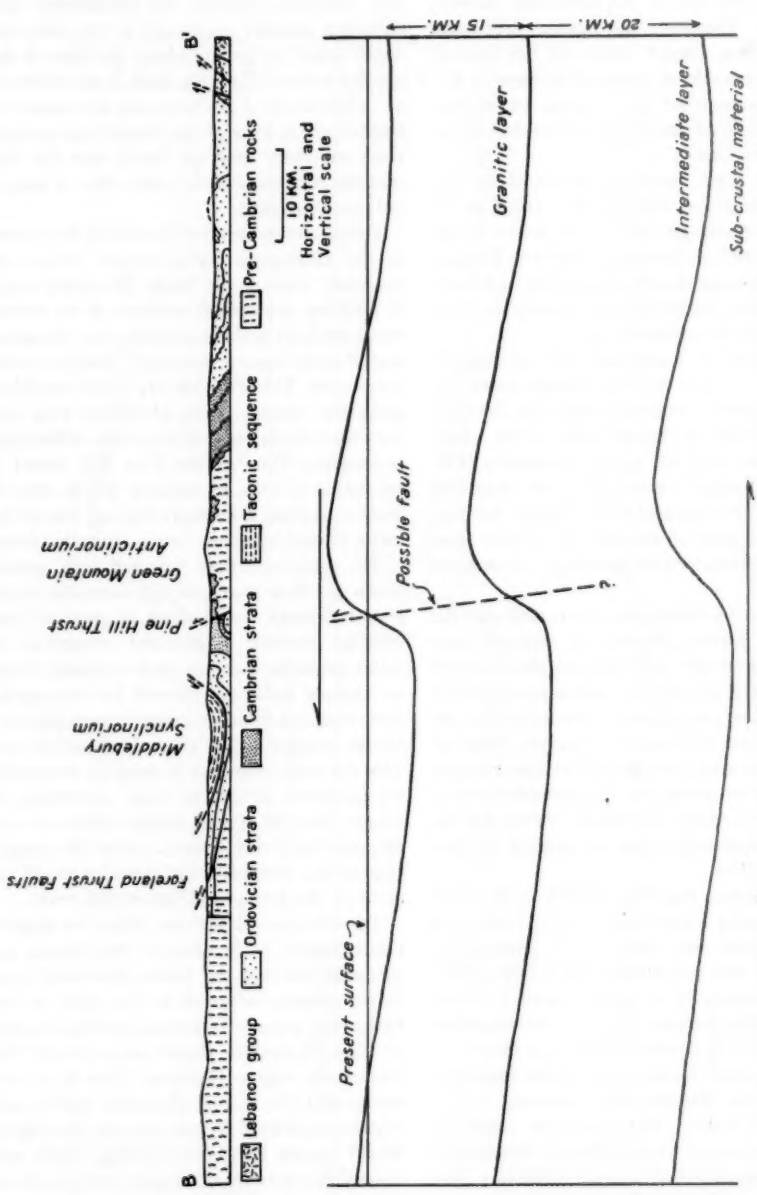


FIGURE 6.—COMPARISON OF GEOLOGIC SECTION WITH HYPOTHETICAL CONFIGURATION OF PRECAMBRIAN BASEMENT IN GREEN MOUNTAIN-CHAMPLAIN VALLEY REGION

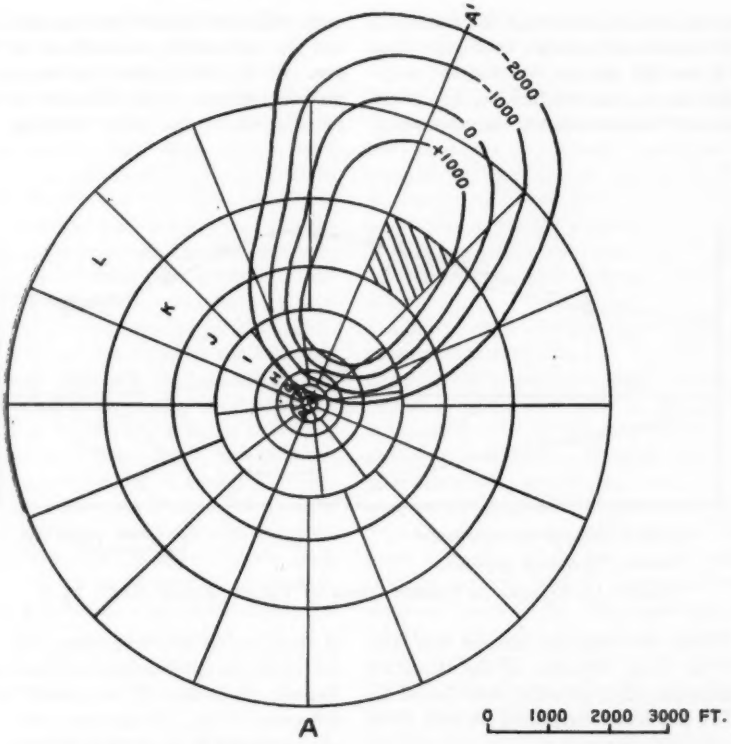


FIGURE 7.—CALCULATION OF THE GRAVITATIONAL ATTRACTION OF BODIES OF FINITE EXTENT

Calculation of the Gravitational Attraction of Bodies of Finite Extent

The vertical component of the gravitational attraction of bodies of finite extent was computed by means of the same template that was used to determine terrain corrections. With the equation that gives the attraction of a vertical cylinder at a point on its axis (Heiland, 1946, p. 157), we may compute the vertical component of the gravitational attraction of a volume of rock whose area corresponds to a compartment of the template, and whose thickness and depth are arbitrary (refer to shaded area in Fig. 7).

If we wish to obtain the vertical component

of the gravitational attraction of the dome outlined by structural contours in Figure 7 at Point A, we first use horizontal slabs to approximate the structural feature. We then count the number of compartments in each zone which occur within the 1000-foot contour line. The sum of the values obtained for the attraction of each compartment gives the attraction of the slab. This procedure is repeated for each given depth and thickness used.

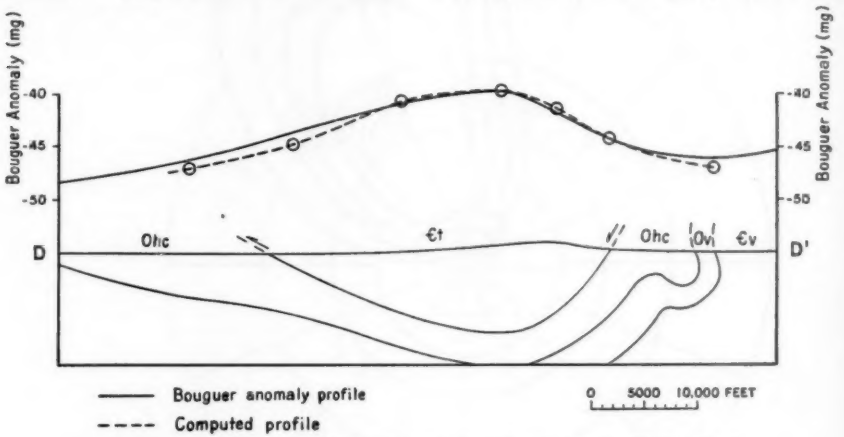


FIGURE 8.—CALCULATED CONFIGURATION OF TACONIC KLIPPE (D-D', Pl. 1)

Gravity High of Taconic Mountains

Within the broad gravity low of the Champlain lowland, a local gravity high occurs over the central and eastern part of the Taconic allochthone. The anomaly results from a density contrast between slate of the Taconic sequence (average density—2.82) and Ordovician limestone and slate (average density—2.72) of the Valley sequence. The anomaly occurs only over the eastern portion of the thrust sheet, because Bomoseen grit of low density (2.72 ?) is present in the western part of the area. The small minimum which occurs between the Taconic high and the Green Mountain high near the northern limit of the thrust is absent in the south where the slates of the Taconic range approach the Green Mountain front. In the latter area, the Taconic high merges with the Green Mountain high.

Although the gravity high appears to be definitely related to the Taconic klippe, the contours which outline the high are very irregular and are not entirely confined to the Taconic area (Pl. 1). This is probably because of the presence of some thick dolomite in the surrounding Ordovician valley sequence. In spite of these limitations, an attempt was made to determine the approximate configuration of the Taconic allochthone. If we assume a density difference of 0.1, calculations show that the klippe attains a maximum thickness of 8600 feet in the eastern part of the thrust sheet, and the thrust plane has steeper dips on the east (Fig. 8). Because very little evidence is available concerning the dip of the thrust plane, this conclusion cannot be checked against geologic data.

The rocks in both the Middlebury synclinorium and the Taconic allochthone appear to attain their maximum thickness in the eastern part of the Champlain lowland. This suggests that the Taconic thrust was emplaced prior to the close of the orogeny in the Green Mountain region, and that the rocks of the Taconic thrust sheet were folded in a manner similar to those of the autochthonous series in the Middlebury synclinorium.

Eastern Vermont and Western New Hampshire

To the east of the Green Mountain anticlinorium, a gravity high trends north-northwestward through the area (Pl. 1). This high coincides with the outcrop of the Pinney

Hollow, Ottauquechee, Bethel, and Missisquoi formations, which have an average density of 2.82. Formations immediately to the east and west of this trend have a consistently lower density.

Regional gravity data show that a local gravity low is associated with the Pomfret dome, and that a low is possibly associated with the Strafford dome. In addition, definite indications are present that a gravity minimum occurs in the northeastern portion of the Woodstock quadrangle. In western New Hampshire, relatively large negative gravity anomalies are found over the Lebanon and Mascoma domes. These gravity anomalies will be considered in detail in the section dealing with the interpretation of complete Bouguer anomalies.

The dominant regional gravity trends in this region strike north-south in eastern Vermont, but gradually attain a more northeasterly trend in the Mt. Cube quadrangle. Bouguer anomalies descend from a maximum of −3 mg. along the axis of the Green Mountain anticlinorium to a broad gravity low in eastern Vermont and western New Hampshire. The minimum anomaly in this region, associated with the Lebanon dome, is −50 mg.

Central and Eastern New Hampshire

The northeasterly gravity trends in the Mt. Cube quadrangle swing around to become a dominant east-west trend in central and eastern New Hampshire. Bouguer anomalies decrease from a maximum value of −23 mg. in the southwestern portion of the Cardigan quadrangle to a minimum value of −57 mg. on the stock of Conway granite in the northeastern part of the Plymouth quadrangle. The regional decrease of the anomalies toward the north is evidently associated with the granitic rocks of the White Mountain batholith, the southern part of which extends into the area surveyed.

Because the regional Bouguer anomaly map indicates a close correspondence between certain lithologic units and gravity trends, a residual anomaly map was constructed for the region (Pl. 3a). The regional anomaly in this area decreases northward at a rate of 0.76 mg. per mile. The high density (2.82) Littleton formation is nearly always associated with gravitationally high areas. A striking example of this is found in an arcuate gravity high which parallels the outcrop of the Littleton formation in the Holderness, Plymouth, and Mt. Chocorua quadrangles. Gravitationally low areas are consistently associated with outcrops of the Concord granite (2.67) and Kinsman quartz monzonite (2.70). The close association of gravity anomalies with an area of Kinsman quartz monzonite may be seen in the Holderness and Winnipesaukee quadrangles.

A gravity high over the western part of the Ossipee Mountain ring-dike complex appears to be associated with the Moat volcanics. Although some rhyolites occur in the Moat volcanics, most of the rocks are high density (2.90) basalts and andesites (Kingsley, 1931). Unfortunately, only one gravity station could be accurately located on this lithologic unit.

In the eastern part of the Ossipee Mountains, a gravity profile was located over the stock of Conway granite. Anomalies increase westward toward the high associated with the Moat volcanics. Although the eastern lowlands which border the Ossipee Mountains are covered with glacial drift, the anomaly map shows that the Concord granite (2.67) probably extends into this area. However, since the gravity profile shows a gradual westward increase in anomalies from the Concord to the Conway granite, any small anomaly associated with the granite contact is apparently obscured by the western gravity high. The anomalies decrease much more rapidly eastward from the gravity high than westward. This is because granites lie to the east, whereas quartz diorite is present toward the west.

In an attempt to obtain the approximate depth of the Littleton schist in central and eastern New Hampshire, computations were carried out for five areas.

(1) The difference in residual anomaly over the Littleton formation (2.82) at the border of the Rumney and Cardigan quadrangles is +6 mg. The average density of the surrounding rocks, which are the Bethlehem gneiss and the Kinsman quartz monzonite, is 2.73. Because the contours are essentially concentric, the depth of the Littleton formation in this area was computed by means of the equation which

gives the gravitational attraction at a point located on the axis of a vertical cylinder. If the depth to the top of the cylinder is zero, we have:

$$\Delta g = 2\pi k\rho[L + r - \sqrt{L^2 + r^2}],$$

where Δg is the gravitational attraction, k is the gravitational constant, ρ is the density contrast, L is the length of the cylinder, and r is the radius of the cylinder. In the present case, we have a density contrast of 0.09 and a radius of approximately 15,000 feet. Therefore the depth of the Littleton schist in this area is approximately 6600 feet.

(2) The difference in residual anomaly over the Concord granite (2.67) in the Rumney and Plymouth quadrangles is −6 mg. The surrounding rocks belong to the Littleton formation (2.82). Because the contours are elongated parallel to the intrusion, a rectangular prism of granite 21,300 feet wide and 49,000 feet long was chosen for depth computations. The density contrast (0.15) was found to persist to a depth of 3700 feet.

(3) The difference in anomaly over the Kinsman quartz monzonite (2.70) and Winnipesaukee quartz diorite (2.73) in the Holderness and Winnipesaukee quadrangles is −10 mg. If an average density of 2.72 is taken for the rocks, the density contrast with the Littleton formation is 0.10. The depth of a vertical cylinder of 15,000 feet radius was found to be 8700 feet.

(4) The difference in residual anomaly over the Concord granite (2.67 ?) in the Wolfeboro quadrangle is −6 mg. The Littleton formation (2.82) is the principal lithologic unit into which the granite is intrusive. Because of the approximately circular shape of the anomaly contours, a vertical cylinder of 15,000 feet radius and density contrast 0.15 was chosen for the depth computation. The density contrast persists to a depth of 3500 feet.

(5) The Littleton formation was found to be approximately 4000 feet deep near the southwest contact of the stock of Conway granite in the Plymouth quadrangle. This area will be considered in more detail in a later section.

These calculations indicate that the schists do not continue downward to great depths as is shown in Plate 1, but are cut off at relatively shallow depths.

Although the relatively shallow depth of the Littleton formation east of the Bronson Hill anticline may be attributed to the gradation of schist into relatively low density gneiss of a higher metamorphic facies, there is no surface evidence in this region that such a process occurs on a large scale at relatively shallow depths. It is also possible that the intrusives of the New Hampshire magma series become wider at shallow depths and cut off the Littleton formation. In the Rumney and Cardigan quadrangles, the Littleton formation may be cut off at depth by the Bethlehem gneiss which dips eastward at angles ranging from 50° to 60°. The dips must decrease downward to yield a depth of 6600 feet for the schist in the area.

Another possibility is that rocks of the Oliverian magma series are present beneath the region. These rocks could be present at relatively shallow depths throughout the entire area, but were domed upward and are intersected by the present erosion surface only along the axis of the Bronson Hill anticline.

One of the problems of New Hampshire geology is whether a master reservoir of the White Mountain magma series underlies the ring-dike complexes, or whether the complexes represent the outcrop area of deep, cylindrical intrusions which descend to a depth of 15 or 35 km. (Billings, 1942). The Littleton schist is deepest in those regions which are farthest removed from areas where the White Mountain magma series outcrops. Whereas the depth of the formation ranges from 3500 to 4000 feet in the Plymouth and Wolfeboro quadrangles, the depth increases to 6600 feet in the northern part of the Cardigan quadrangle, and to 8700 feet in the Holderness quadrangle. If this variation in depth is attributed to a master reservoir of the White Mountain magma series beneath the region, the ring-dike complexes probably originated as cupolas of magma which approached relatively close to the surface during the Mississippian period.

Analysis of Complete Bouguer Anomalies

Residual Anomalies of Eastern Vermont

General Discussion.—The eastern Vermont area is situated between the north-south trending Green Mountain gravity high, and the

north-northeast trending western New Hampshire gravity low. Because the regional gravity field, which decreases eastward at a rate of 1.1 mg. per mile, interfered with the interpretation of local anomalies, it was removed. To obtain the greatest accuracy possible, all of the stations within the area were corrected for terrain effects, and the average density of the rocks in the region was used in the Bouguer correction. Consequently, most of the anomalies are probably accurate to 1 mg. Both of the predominant lithologic units in the area, the Waits River and the Gile Mountain formations, have an average density of 2.76.

An arcuate gravity high extends from the north-central part of the Strafford quadrangle to the west-central part of the Hanover quadrangle (Pl. 3b). This trend appears to coincide with the outcrop of the Standing Pond amphibolite (density, 3.02). The largest anomalies are associated with either the thickest section of the amphibolite, or with highly deformed areas, where the amphibolite is probably concentrated as a result of intense folding.

A gravity high in the southeastern part of the Strafford quadrangle appears to be related to another high in the east-central part of the Hanover quadrangle. Both are associated with areas in which the Post Pond volcanics (2.85) occur. Although a small gravity high joins the two anomaly fields, the trend is obscured by the presence of a large gravity minimum in the area associated with the Lebanon granite.

Small gravity lows are related to outcrop areas of the Barnard gneiss (2.70) in the central part of the Woodstock quadrangle.

The dominant feature of the residual gravity map is the presence of three gravity minima. One of these is over the Strafford dome, another is over the Pomfret dome, and the third is north of Woodstock, Vermont. As no appreciable density differences exist between surface rocks in this region, the anomalies indicate that a mass of rock of lower density than the surrounding formations is present at depth beneath these areas.

To interpret these anomalies, it is necessary to make an assumption as to the density difference which produces the anomalies. This was done by taking into consideration the density of the core rocks in the Lebanon and Mascoma domes of western New Hampshire, and the Chester dome, located south of the Woodstock quadrangle. The following average densities are taken from Table 1:

	Average Density
Lebanon granite	2.66
Mascoma group	2.68
Gneiss in Chester dome	2.69

On the basis of these data, the anomalous mass beneath the domes of eastern Vermont is assumed to possess a density of 2.68. Therefore, the density contrast is 0.08.

Strafford dome.—The residual anomaly map shows that a gravity low occurs over the Strafford dome, a north-south trending, doubly plunging anticline in the south-central part of the Strafford quadrangle (Doll, 1944). A gravity profile drawn perpendicular to the axis of the anomaly pattern, and across the center of the gravity minimum, shows that an average difference in anomaly of −3 mg. is present (Fig. 9). As the gravity profile is essentially symmetrical, the interpretation is based on simple, symmetrical geometrical forms. Because of the many uncertainties in the interpretation of a wholly subsurface anomalous mass, additional refinement would be unwarranted. The anomaly map shows that the gravity contours have a roughly elliptical shape; consequently, the anomaly curve of a finite horizontal cylinder is compared with an average gravity profile across the dome. Calculations indicate that a cylinder which has an axial depth of 8500 feet and a radius of 5800 feet approximates the mass distribution at depth. The deviation of the residual anomaly profile from the computed curve (Fig. 10) at the northern end of the Strafford dome is due to the gravity high which is probably associated with the Standing Pond amphibolite; the deviation at the southern end of the profile results from the gravity low associated with the Pomfret dome.

If the structure of the Strafford dome is similar to that of the Chester dome, the above calculations show that the depth to the Precambrian gneiss should be of the order of 2500 feet. Thompson (1950) reports that the thickness of the Cambro-Ordovician section to the base of the Waits River formation is as little

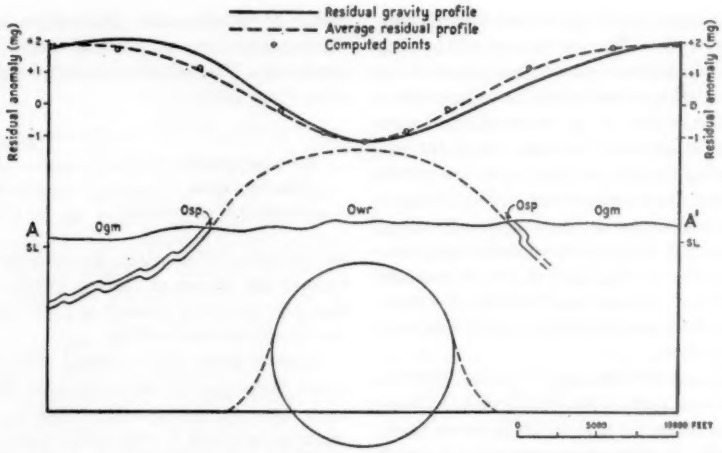


FIGURE 9.—CALCULATED ANOMALOUS MASS ASSOCIATED WITH THE STRAFFORD DOME (A-A′, Pl. 3b)

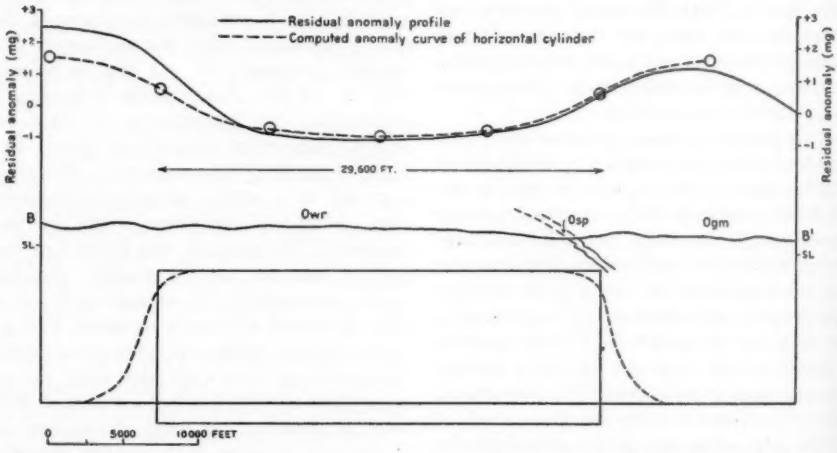


FIGURE 10.—CALCULATED ANOMALOUS MASS ASSOCIATED WITH THE STRAFFORD DOME (B-B′, Pl. 3b)

as 1500 feet at the north end and on the flanks of the Chester dome. Therefore, the calculated depth to the anomalous mass is not inconsistent with this hypothesis. On the other hand, the Strafford dome could have resulted from an igneous intrusion in a manner similar to that of the Lebanon dome.

Pomfret dome.—The Pomfret dome in the northwestern part of the Hanover quadrangle is a north-south trending, doubly plunging anticline. The foliation of the rocks over the dome dips radially outward at angles ranging from 30° to 50°. Although a distinct gravity minimum coincides with the outcrop of the Waits River formation in the center of the dome, a broad negative anomaly extends toward the southeast. In order to account for this anomaly distribution, it is assumed that the mass deficiency may be approximated by two coaxially placed vertical cylinders superimposed on a finite horizontal cylinder. Although a gravity profile drawn perpendicular to the trend of the anomaly pattern and across the center of the gravity minimum shows that a maximum dif-

ference in anomaly of −4 mg. is present, several stations have differences in anomalies of −3 mg. in the southeastern extension of the anomaly field. A horizontal cylinder 39,000 feet long is used to approximate a northwest-southeast trending mass which is believed to produce the broad anomaly field (Fig. 11). The axis of this cylinder is placed at a depth of 8300 feet, which is equal to the half-width of the profile drawn across the center of the negative anomaly field (C-C′, Pl. 3b). Calculations show that a radius of 5300 feet is necessary for such a mass to produce a maximum anomaly of −3 mg. The configuration of mass which produces an anomaly curve that best fits the observed data over the gravity minimum is shown in Figure 12. Calculations indicate that the depth to the top of the anomalous mass is 1000 feet, and the depth to the base is approximately 13,000 feet. The gravitational attraction of cylindrical discs was obtained by use of the solid angle chart of Nettleton (1942).

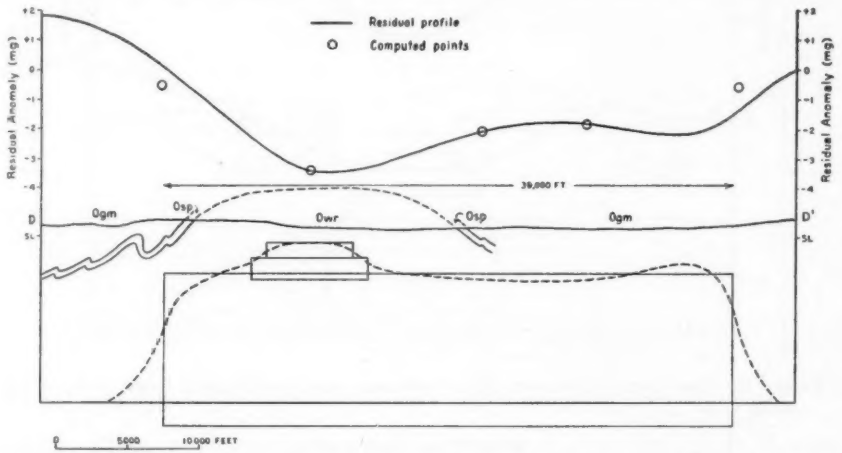


FIGURE 11.—CALCULATED ANOMALOUS MASS PARALLEL TO THE AXIS OF THE POMFRET LOW (D-D′, Pl. 3b)

Although the depth to the top of the mass deficiency is of the order of magnitude of the thinnest Cambro-Ordovician sections below the Waits River formation around the Chester dome, the interpretation would have to be modified to allow for the greater density of the Pinney Hollow-Missisquoi sequence. Therefore, if the Waits River formation is relatively thin over the Pomfret dome, the assumed density contrast is too small, and the mass deficiency is located at a greater depth than that calculated. On the other hand, if the Strafford and Pomfret domes resulted from an igneous intrusion in a manner similar to that of the Lebanon dome, the thickness of the Cambro-Ordovician section would not have to be taken into consideration.

The regional gravity low in eastern Vermont suggests thickening of the crust in this region. If the domes originated through differential vertical movements of the Precambrian basement, plastic thickening of the crust is indicated. Under horizontal compressive forces, downward movement of lower crustal materials might occur simultaneously with upward movements in the higher portions of the granitic layer. The fact that the high-grade metamorphic zones of eastern Vermont are associated with regions of domal uplift suggests that this area was subjected to the maximum intensity of the orogeny.

The negative anomalies associated with the domes of eastern Vermont does not support the theory of their origin presented by White and Jahns (1950). These authors believe that the doming resulted from the upward migration of mobile calcareous rocks of the Waits River formation under the action of horizontal com-

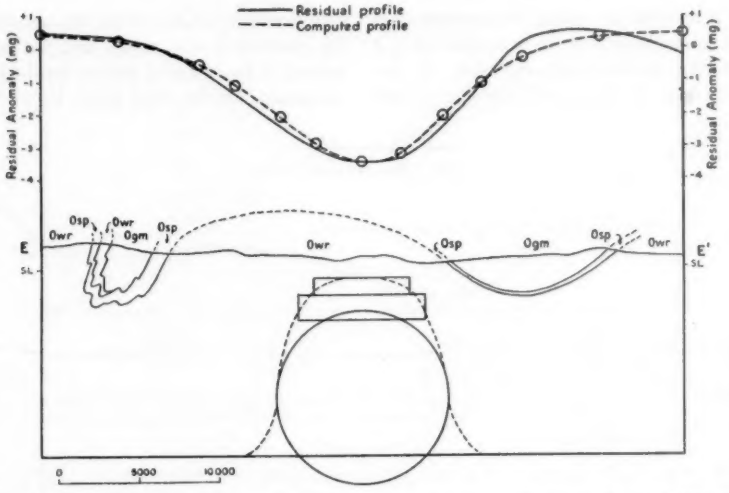


FIGURE 12.—CALCULATED ANOMALOUS MASS ASSOCIATED WITH THE POMFRET DOME (E-E', Pl. 3b)

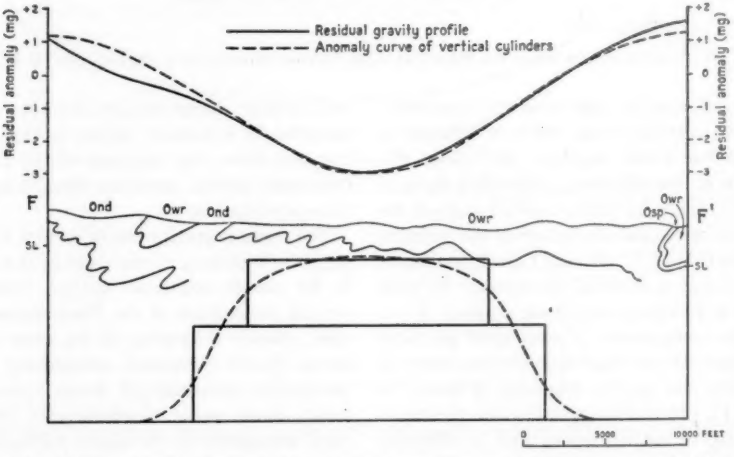


FIGURE 13.—CALCULATED ANOMALOUS MASS ASSOCIATED WITH THE WOODSTOCK LOW (F-F', Pl. 3b)

pressive forces. If this theory were correct, no anomaly should result, since only high density rocks would be involved in the uplift.

Woodstock low.—A broad gravity low in the northeast and east-central parts of the Woodstock quadrangle is distributed over a broad area, but the minimum anomalies occur in the southern part of the anomaly field (Pl. 3b). An east-west gravity profile across the center of the minimum shows a difference in anomaly of approximately −4.5 mg. (Fig. 13). Because the residual anomaly contour over the minimum is circular, an anomaly curve was calculated for a cylindrical distribution of mass. Although the anomaly map shows that the distribution of mass probably extends northward, the simplified calculations are sufficient to indicate the order of magnitude of the depth to the top and bottom of the anomalous mass.

The cylindrical configuration whose anomaly curve best fitted the observed data consists of two coaxial cylinders. Calculations indicate that the depth to the top of the anomalous mass is approximately 2000 feet, and the depth

to the base of the configuration is 12,000 feet. As this depth is of the same order of magnitude as that calculated for the Strafford and Pomfret domes, the cores of these structural features because this region occupies a broad gravity low between the Green Mountain high on the west and a high in the Cardigan and Rumney quadrangles on the east. As no pronounced

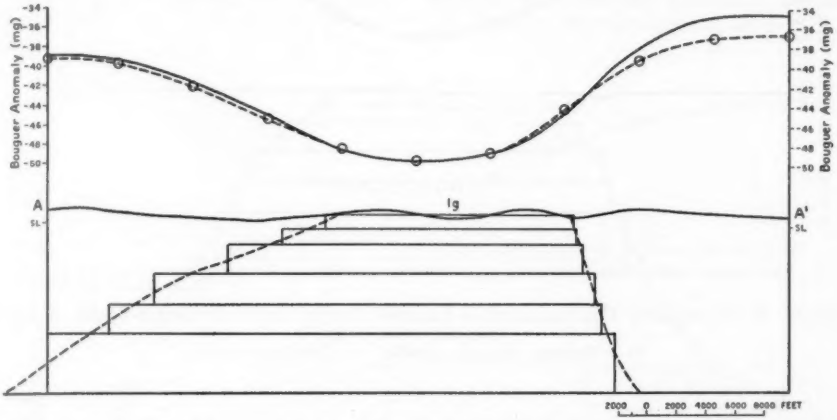


FIGURE 14.—CALCULATED CONFIGURATION OF LEBANON GRANITE ALONG SECTION A-A′ (Pl. 2) ACROSS THE LEBANON DOME
(——Bouguer anomaly profile - - -Calculated profile)

may terminate in a relatively low density rock which occurs at depth throughout the region. However, whether the cores originated as basement uplifts or as igneous intrusions is still indeterminant. This subject will receive further consideration during the discussion of the Mascoma dome.

Although the anomaly in the Woodstock area does not overlie a dome which has been outlined by marker formations, the strike and dip of the foliation in the Waits River formation indicates a north-south anticlinal trend in the region. Moreover, there is a tendency for the foliation to assume east-west strikes and southerly dips south of Woodstock, Vermont. In other words, the anticlinal axis plunges south in the region south of the minimum anomaly. Since detailed work has not been done in the northern part of the Woodstock quadrangle, it is not known whether the anticline plunges northward in the area.

Complete Bouguer Anomalies of Western New Hampshire

General discussion.—Although all gravity stations in this area were corrected for terrain effects, no residual anomalies were computed regional anomaly gradient occurs in this region, the anomaly contours are not appreciably disturbed.

Large gravity lows are associated with the Lebanon dome in the Hanover and Mascoma quadrangles, and with the Mascoma dome in the central part of the Mascoma quadrangle (Pl. 2). The dominant gravity highs in the region are found over areas of the Post Pond volcanics. A gravity low is apparently associated with the Smarts Mountain pluton in the southeast part of the Mt. Cube quadrangle, but, unfortunately, only a few stations could be occupied in this area.

Although a large negative anomaly appears to be associated with the Fairlee quartz monzonite, Diment (personal communication) reports that the trend continues to the northeast through the Woodsville quadrangle.

Lebanon dome.—The anomaly map shows that a large gravity low is associated with the Lebanon complex. An average difference in anomaly of −13 mg. results from an intrusion of low density Lebanon granite (2.66) into relatively high density rocks of the Orfordville formation (2.75) and Post Pond volcanics (2.85). In addition, the Lebanon border gneiss has a density of 2.79. As the average density

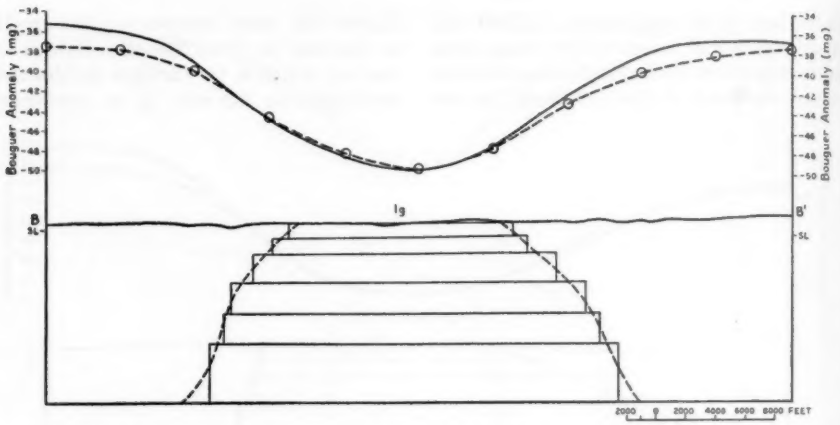


FIGURE 15.—CALCULATED CONFIGURATION OF LEBANON GRANITE ALONG SECTION B-B′ (Pl. 2) ACROSS THE LEBANON DOME
(——Bouguer anomaly profile - - -Calculated profile)

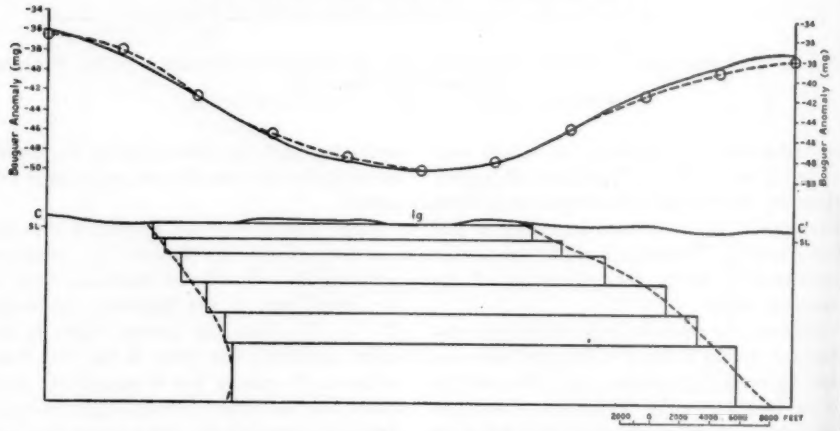


FIGURE 16.—CALCULATED CONFIGURATION OF LEBANON GRANITE ALONG SECTION C-C′ (Pl. 2) ACROSS THE LEBANON DOME
(——Bouguer anomaly profile - - -Calculated profile)

of the country rock is taken as 2.79, a density difference of 0.13 is obtained. For purposes of computation, the border gneiss is considered as country rock.

To determine the general shape of the granite pluton and the depth to which the density difference persists, computed anomaly curves are compared with complete Bouguer anomaly profiles across the intrusion (Figs. 14, 15, 16). Calculated profiles indicate that, although the dips of the contact near the surface are essentially parallel to the foliation of the rocks on the flanks of the dome, the dips appear to increase at depth. Moreover, the calculations indicate that the density contrast extends to a depth of approximately 12,000 feet and confirm geologic data that the Lebanon dome is overturned toward the south-southeast.

Mascoma dome.—The Bouguer anomaly map shows that a difference in anomaly of approximately −10.5 mg. is associated with the rocks of the Oliverian magma series in the Mascoma dome (Pl. 2). The Mascoma group consists of granite, quartz monzonite, granodio-

rite, and quartz diorite. Density values, obtained for all the units except the quartz diorite, are as follows:

	Average Density
granite and quartz monzonite	2.68
granodiorite	2.71

crease at depth and that the density contrast extends to a depth of approximately 16,000 feet (Fig. 17). As in the case of the Lebanon dome, the dips of the contact near the surface are found to be essentially parallel to the foliation of the rocks on the flanks of the dome.

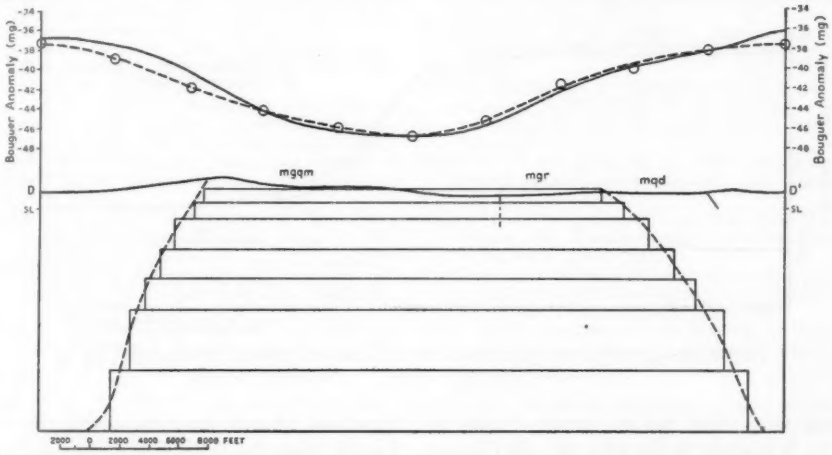


FIGURE 17.—CALCULATED CONFIGURATION OF MASCOMA GROUP ALONG SECTION D-D' (Pl. 2)
(——Bouguer anomaly profile - - -Calculated profile)

The density of the rocks surrounding the dome are given below.

	Average Density
Littleton formation (Medium-grade)	2.76
Orfordville formation	2.75
Bethlehem gneiss	2.75
Ammonoosuc volcanics	2.96
Clough formation	2.68

Because the Ammonoosuc volcanics are relatively thin along the flanks of the Mascoma dome, they disturb the gravity contours only in the east-central part of the Mascoma quadrangle. Inasmuch as the largest area of quartz diorite is in the eastern part of the dome, it is assumed to have a density equal to that of the Bethlehem gneiss (2.75). For purposes of computation, the quartz diorite is considered as country rock. To calculate the effect of the granodiorite, a density difference of 0.04 is used, and a density difference of 0.07 is used to calculate the effect of the granite or quartz monzonite. The average density of the surrounding rocks is assumed to be 2.75. Calculations indicate that the dips of the contact in-

These results show that relatively high density schists and gneisses surround a solid mass of low density rock of the Oliverian magma series to a depth of 16,000 feet. Therefore, the data indicate that the Mascoma dome is not the result of a phacolithic intrusion. However, either a thick laccolithic intrusion or a thick, sheetlike intrusion which has been domed upward could explain the gravity data.

On the other hand, the fact that the estimates of depth of the Lebanon, Pomfret, Strafford, and Mascoma domes are of the same order of magnitude suggests that the core gneiss in the Mascoma dome may be an uplift of a low density basement complex which underlies the high density metamorphic rocks in the region. This interpretation would support the theory that the Oliverian magma series resulted from metamorphism and uplift of early Paleozoic volcanics or graywackes. This theory would also explain the fact that the Ammonoosuc volcanics immediately overlie the core rocks throughout the length of the Bronson Hill anticline.

Stock of Conway Granite

Eight stations were occupied in a 10 mile profile along the Mad River in the Plymouth quadrangle. A station interval ranging from 1 to 1½ miles was used. Only two lithologic units are included in the profile—the Littleton formation, which has an average density of 2.87 (9 samples) in the area; and the Conway granite, which has an average density of 2.62. Consequently, a 0.25 density difference is present to produce the difference in anomaly of −11 mg. The regional Bougeur anomalies were corrected to take into consideration the average density of the surface rocks in the region, and all stations were corrected for terrain effects.

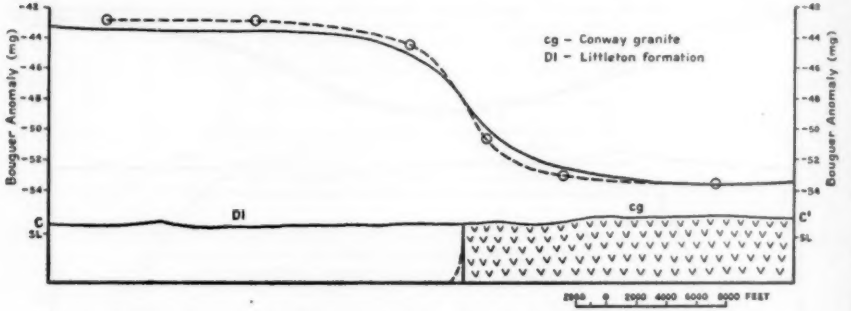


FIGURE 18.—CALCULATED CONFIGURATION OF CONWAY GRANITE STOCK (C-C', Pl. 1)
(——Bouguer anomaly profile - - -Computed profile)

The Conway granite stock is an elliptically shaped intrusion approximately 6.5 miles long and 5 miles wide (Pl. 1). Moke (1945) reports that, although the contact could be seen at only a few points, one of these was along the Mad River. In some areas, the dip is less than 60°, whereas in others, such as the Mad River valley, the manner in which the contact disregards topography suggests that it is essentially vertical, which is supported by the fact that approximately half the maximum difference in anomaly occurs at the point of contact (Fig. 18).

The assumption was made that the contact of the stock is vertical in all areas, and an anomaly curve was calculated along the line of the profile. The computed curve shows that this assumption is not greatly in error, and that the density difference exists to a depth of approximately 4000 feet.

CONCLUSIONS

Gravity anomalies in the Green Mountain-Champlain Valley region may be related to warping or to high-angle thrust faulting of the crust during the Taconic orogeny. Both geologic and gravimetric data are consistent with this hypothesis. Gravity data also indicate that the Taconic klippe has the same general structural configuration as the Middlebury synclinorium. Therefore, it seems probable that the rocks of the thrust sheet were folded concurrently with the rocks of the synclinorium. In addition, a series of domes which originated in eastern Vermont as a result of this orogeny are related to gravity minima. These anomalies must result from a mass of relatively low density rock beneath the high density metamorphic rocks at the surface. Calculations show that these core rocks persist to depths of the same order of magnitude as those of the Lebanon and Mascoma domes. Consequently, the core rocks may be either intrusions or uplifts of a low density rock which occurs at depth throughout the region. In central New Hampshire, gravity data indicate that the Littleton formation is probably cut off at relatively shallow depths by one or more of the magma series which occur in the region.

REFERENCES CITED

Bickel, H. C. (1948) *A note on terrain corrections*, Geophys., vol. 13, p. 255–258.

Billings, M. P. (1928) *Petrology of the North Conway quadrangle in the White Mountains of New Hampshire*, Am. Acad. Arts. and Sci., Pr., vol. 63, p. 67–137.

—— (1942) *Geology of the central area of the Ossipee*

Mountains, New Hampshire, earthquakes, Seismol. Soc. Am. Bull., vol. 32, p. 83–92.

—— (1945) *Mechanics of igneous intrusion in New Hampshire*, Am. Jour. Sci., 5th ser., vol. 243-A, p. 40–68.

Cady, W. M. (1945) *Stratigraphy and structure of west-central Vermont*, Geol. Soc. Am., Bull., vol. 56, p. 515–588.

Chang, P. H. (1950) *Structure and metamorphism of the Bridgewater-Woodstock area, Vermont*, Ph.D. Thesis, Harvard Univ.

Chapman, Carleton A. (1939) *Geology of the Mascoma quadrangle, New Hampshire*, Geol. Soc. Am., Bull., vol. 50, p. 127–180.

Doll, C. G. (1944) *A preliminary report on the geology of the Strafford quadrangle, Vermont*, Vt. State Geol., 24th Rpt., 1943–44, p. 14–28.

Fowler, Phillip (1949) *Stratigraphy and structure of the Castleton area, Vermont*, Harvard Univ., Ph.D. Thesis.

Glennie, E. A. (1933) *Crustal warpings*, Mon. Not. Royal Astron. Soc., Geophys. Supp., no. 3, p. 170–176.

Hadley, J. B. (1942) *Stratigraphy, structure, and petrology of the Mt. Cube area, New Hampshire*, Geol. Soc. Am., Bull., vol. 53, p. 113–176.

Hammer, Sigmund (1939) *Terrain corrections for gravimeter stations*, Geophys., vol. 4, p. 184–194.

Heiland, C. A. (1946) *Geophysical exploration*, New York.

Hubbert, M. K. (1948) *A line integral method of computing the gravimetric effects of two-dimensional masses*, Geophys., vol. 13, p. 215–225.

Kaiser, E. P. (1945) *Northern end of the Taconic thrust sheet in western Vermont*, Geol. Soc. Am., Bull., vol. 56, p. 1079–1098.

Kingsley, Louise (1931) *Cauldron subsidence of the Ossipee Mountains*, Am. Jour. Sci., vol. 22, p. 139–168.

Leet, L. D. (1941) *Trial travel times for northeastern America*, Seismol. Soc. Am. Bull., vol. 31, no. 4, p. 325–334.

Longwell, C. R. (1943) *Geological interpretations of gravity anomalies in the southern New England-Hudson Valley region*, Geol. Soc. Am., Bull., vol. 54, p. 555–590.

Moke, C. B. (1945) *Petrology, structure, and metamorphism of the Plymouth quadrangle, New Hampshire*, Harvard Univ., Ph.D. Thesis.

Nettleton, L. L. (1940) *Geophysical prospecting for oil*, McGraw Hill, 444 pages.

—— (1941) *Relation of gravity to structure in the northern Appalachian area*, Geophys., vol. 6, p. 270–286.

—— (1942) *Gravity and magnetic calculations*, Geophys., vol. 7, p. 293–310.

Thompson, J. B., Jr. (1950) *A gneiss dome in southeastern Vermont*, Mass. Inst. Tech., Ph.D. Thesis.

Vening Meinesz, F. A., Umbgrove, J. H. F., and Kuenen, P. H. (1934) *Gravity expeditions at sea* (1923–32), vol. II, Neth. Geod. Comm., 208 pages.

White, W. S. and Jahns, R. H. (1950) *Structure of central and east-central Vermont*, Jour. Geol., vol. 58, p. 179–220.

Wilson, J. H. (1941) *Gravity meter survey of the Wellington field, Larimer County, Colorado*, Geophys., vol. 6, p. 264–269.

Woollard, G. P. (1939) *The geological significance of gravity investigations in Virginia*, Am. Geophys. Union, Tr., 20th ann. mtg., p. 317–323.

—— (1940) *Gravitational determination of deep-seated crustal structure of continental borders*, Am. Geophys. Union, Tr., 21st ann. mtg., p. 808–815.

——(1943) *Geological correlations of areal gravity and magnetic studies in New Jersey and vicinity*, Geol. Soc. Am., Bull., vol. 54, p. 791–818.

—— (1944) *Gravity observations in northern New England*, Am. Geophys. Union, Tr., 25th ann. mtg., p. 254–258.

—— (1948) *Gravity and magnetic investigations in New England*, Am. Geophys. Union, Tr., vol. 29, p. 306–317.

Shell Oil Company, Houston, Texas.

Paper No. 125. Published Under the Auspices of the Committee on Experimental Geology and Geophysics and the Division of Geological Sciences at Harvard University.

Manuscript Received by the Secretary of the Society, July 24, 1951.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 64, PP. 539-546, 2 FIGS., 3 PLS. MAY 1953

CENOZOIC FILL AND EVAPORATE SOLUTION IN THE DELAWARE BASIN, TEXAS AND NEW MEXICO

By V. C. Maley and Roy M. Huffington

Abstract

The Delaware Basin, containing thick deposits of late Permian salt and anhydrite, is one of the major geologic subdivisions of the Permian Basin of West Texas and southeastern New Mexico. At least three extensive accumulations of Cenozoic fill, with a maximum thickness of approximately 1900 feet, are present in the basin. Localization of these major fill deposits is attributed to evaporite solution, particularly of salts. Consequent slump of the Permian Rustler formation, which immediately overlies the evaporite beds, makes it difficult to distinguish true tectonic structures in this formation.

CONTENTS

TEXT

	Page
Introduction	539
Areal extent and age of basin fill	541
Major accumulations of fill	541
Evaporite solution—Chief factor in fill localization	541
Structural features in the Rustler formation	544
Conclusions	545
References cited	545

ILLUSTRATIONS

Figure	Page
1. Index map	540
2. Table of younger stratigraphic units of the Delaware Basin	542
Plate	**Facing page**
1. Isopach map of Cenozoic basin fill and cross sections	539
2. Isopach map of salt in the Castile and Salado formations	542
3. Structural map on the top of the Rustler formation	544

Introduction

The Delaware Basin of West Texas and southeastern New Mexico forms one of the larger geologic subdivisions of the Permian Basin of Texas, New Mexico, Oklahoma, Kansas, and Nebraska. In Texas, the Delaware Basin geographically includes all of Reeves and Loving counties, most of Culberson County, and parts of Winkler, Ward and Pecos counties. Very minor parts of Jeff Davis and Brewster counties can also be included within the basin. To the north, the Delaware Basin extends into New Mexico, where it includes parts of Eddy and Lea counties.

In the subsurface, the Delaware Basin is frequently thought of as that area embraced by the Capitan reef of upper Permian age (Fig. 1). More properly, however, the basin must also include the narrow belt back of the Capitan reef into which the older and deeper-lying sands of the basin extend. In its greater extent, as indicated by the presence of Delaware Mountain sandstones, the Delaware Basin covers an area of approximately 12,000 square miles. Greatest north-south and east-west dimensions are, respectively, about 160 and 100 miles. In outline, the basin is shaped somewhat like an inverted pear with the long axis striking south-southeast.

Physiographic features on and around the periphery of the Delaware Basin include the expressionless Llano Estacado on the northeast and east, the Guadalupe Mountains on the northwest, the Salt Flat bolson and Delaware Mountains on the west, the Apache and Davis mountains on the southwest, and the Glass Mountains on the south.

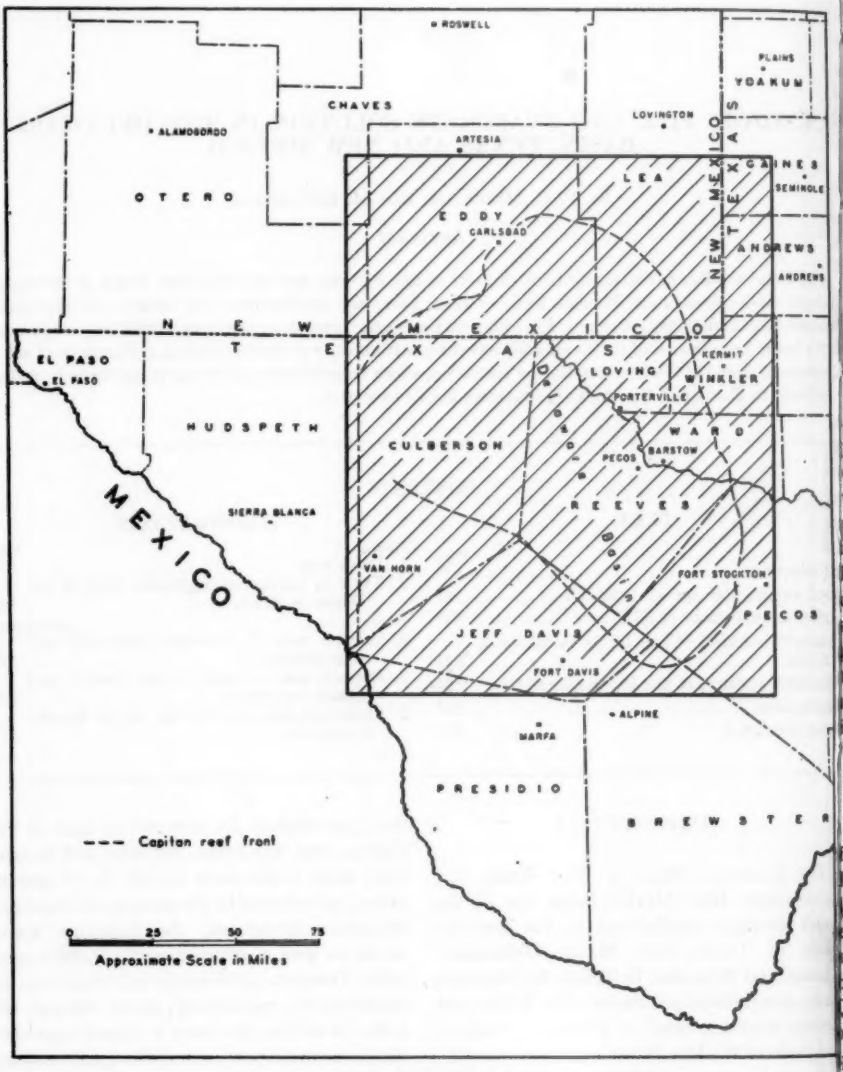


FIGURE 1.—INDEX MAP
Map of part of West Texas and southeastern New Mexico showing location of the Delaware Basin and the area covered by the plates (lined block) in this report.

Topographically, the Delaware Basin consists of a relatively flat to gently sloping and slightly dissected plain covered by alluvium and gravels from the surrounding higher areas. Small outcrops of late Permian, Triassic, and Cretaceous rocks locally protrude through or lie above the gravel plain forming low hills or cuestas. Th throughflowing Pecos River, which enters fro the north and leaves to the southeast, is th main drainage channel of the basin. Othe streams are generally ephemeral.

In addition to satisfying academic curiosity

a knowledge of the extent, thickness, and nature of fill in the Delaware Basin is of economic importance, inasmuch as the blanket of fill either prohibits or greatly restricts the use of surface geologic mapping and seismic prospecting in the constant search for new petroleum reserves.

Because of its relative insignificance in the area discussed, all pre-Cenozoic history of evaporite solution is omitted in this paper. It is recognized, however, that, because of Triassic evaporite solution, a thicker sequence of Triassic rocks was deposited in the small Rustler basin just northeast of the Capitan reef front in Lea County, New Mexico (Pl. 3), than in the immediately adjacent areas.

Sample log information from the Humble Oil and Refining Company files was used in compiling the three accompanying maps. As only driller's logs were available on many of the old wells and samples have never been received on some of the more recent wildcats, the structural and isopach maps are generalized in local areas. Information from future wells is expected to change many of the details, particularly on the structural map of the Rustler formation. In general, however, the configuration and location of several basins and troughs of thick fill accumulations have already been outlined, and it is these that the writers wish to point out.

Areal Extent and Age of Basin Fill

Silts, sands, and gravels which constitute the fill extend over most of the eastern half of the Delaware Basin (Pl. 1). Underlying bedrock is locally exposed along a few low scarps on the eastern side of the basin and on a number of small hills and mesas at the southern end of the basin. Although most of the fill is probably Quaternary in age, it is likely that some of the older deposits are Tertiary, since processes responsible for the accumulation of Quaternary fill must also have been active during much of Tertiary time following the cessation of Cretaceous deposition. It is also expected that, when detailed physiographic and stratigraphic studies of the Cenozoic are carried out in Lea County, New Mexico, and Winkler and Ward Counties, Texas, additional evidence for the suspected Tertiary age of some of the fill will be discovered.

Major Accumulations of Fill

Although fill overlies most of the eastern half of the Delaware Basin, the outcrops give no indication of the considerable difference in thickness of the deposits in this area. Numerous well logs, however, have outlined three separate major accumulations of fill that lie concealed beneath the blanketing surficial layers.

One of these is present in a narrow belt which lies essentially over and just west of the buried Capitan reef on the eastern side of the Delaware Basin (Pl. 1). The belt contains at least three elongate and coalescing lenses of fill, which, when viewed together, are parallel to the Capitan reef front and were obviously influenced by that feature.

The second major accumulation is in a basin which centers about 6 miles north and slightly west of the town of Pecos in Reeves County. Extending outward from this body are several tongues of fill occupying linear depressions whose arrangement is somewhat suggestive of a dendritic stream pattern.

The third body of fill, the smallest of the three, trends northwest and straddles the Texas-New Mexico state line just east of the Pecos River. It is connected to the larger accumulation to the south by thin deposits of fill forming a northward-elongated belt, which suggests that the third lens was deposited in a depression that may once have been tributary to the basin occupied by the larger body. It has evolved to the stage, however, where it can be considered a separate locus of deposition. A similar situation in a more youthful stage may exist just west of Toyah in Reeves County, where another small basin of deposition appears to be developing, gradually detaching itself from the major feature centered just northwest of Pecos.

Evaporite Solution—Chief Factor in Fill Localization

It has been common knowledge for many years among Permian Basin geologists that great amounts of evaporites, particularly salts, have been locally removed from the geologic section in the Delaware Basin as a result of solution by downward-percolating surface

PERIOD	SERIES	GROUP	FORMATION	LITHOLOGY	THICKNESS IN FEET
QUATERNARY & TERTIARY				Basin fill of boulders, gravel, sand, silt, and clay.	0 - 1900 ±
CRETACEOUS				Limestones, fine- to coarse-grained sandstones, and fine conglomerates.	0 - 600 ±
TRIASSIC				Red to maroon shales and red sandstones.	0 - 1400 ±
PERMIAN	OCHOA		DEWEY LAKE	Very fine-grained red to orange-red sandstones and siltstones.	0 - 600 ±
			RUSTLER	Dolomite, anhydrite, red to yellow fine-grained sandstone, and some conglomerate and salt.	0 - 550 ±
			SALADO	Salt, anhydrite, and some dolomite.	0 - 4200 ±
			CASTILE	Alternating laminae of anhydrite and limestone, massive anhydrite, salt, and some limestone.	
	GUADALUPE	DELAWARE MOUNTAIN	BELL CANYON	Buff to tan, fine-grained sandstones and siltstones and dark brown to black limestones.	700 - 1200 ±
			CHERRY CANYON	Buff to tan, fine-grained sandstones and siltstones and dark brown to black limestones.	1000 ±
			BRUSHY CANYON	Mostly medium- to coarse-grained sandstones.	0 - 1000 ±

FIGURE 2.—TABLE OF YOUNGER STRATIGRAPHIC UNITS OF THE DELAWARE BASIN

waters (Adams, 1936; 1944; Horberg, 1949; and others). By showing the coincidence in position of major bodies of fill with areas where extreme amounts of salt solution have taken place, one demonstrates that solution is the chief agent in localization of the thick fill deposits.

In order to show the amount of salt that has been dissolved out of the thick late Permian evaporite series in the Delaware Basin, an isopach map has been constructed on the total amount of salt present in the combined Castile and Salado formations (Pl. 2). These two formations, composed of anhydrite and salt with minor amounts of limestone and dolomite, contain essentially all of the salt in the late Permian formations of the area (Fig. 2). The overlying and relatively thin Rustler formation contains some salt in the northeastern part of the Delaware Basin, but the amount is small enough to be excluded from this isopach map without affecting its value.

Upon comparing the fill and salt isopach maps (Pls. 1, 2), which were contoured independently of each other, it is immediately evident that the belt of thick fill along the eastern edge of the Delaware Basin overlies an area in which salt beds in the Castile and Salado formations are very thin or entirely missing; the three subsidiary basins of fill approximately delineate the areas of no salt in this narrow belt paralleling the Capitan reef front. Similarly, the basin of thick fill that straddles the Texas-New Mexico state line just east of the Pecos River overlies another area in which the total salt section is very thin.

A few miles west of the belt of fill along the eastern edge of the Delaware Basin is a northward-trending linear area overlain by very little fill and containing several outcrops of bedrock. Plate 2 shows that this area, in which basin fill is thin to missing, essentially overlies a belt of maximum salt thickness.

Although the amount of salt solution thus appears to have been a major factor in determining where the thickest fill was deposited, it was not the only factor. Of lesser importance, though of consequence, was the removal of anhydrite by solution. The large basin of fill centered just north of the town of Pecos illustrates this point. Whereas the eastern edge of the basin abuts against the northward-trending belt of thick salt deposits, the western edge and side are located in a region from which all salt has been removed. Thus subdendritic topographic irregularities on the bottom of the basin in the western half must largely be ascribed to solution of anhydrite and gypsum.

It has been suggested that areas of deep fill may represent the former course of the Pecos River. The lack of influence of the river on the position of major accumulations of fill, however, is quite apparent on Plate 1. It is also clear that the fill does not represent the location of channels or valleys cut by one or more prehistoric rivers which are no longer present; according to this concept, the thickness of fill should represent the depth of the former stream valley.

The subdendritic pattern formed by the arms of fill extending out from the major accumulation which centers a few miles north of the town of Pecos does suggest to the writers, however, that some of the present ephemeral surface drainage has left its imprint upon the underlying evaporite deposits. The greater the amount of unsaturated surface water that passes through the evaporite beds, the greater the amount of solution that takes place locally. Hence the degree of solution should be accentuated along the stream courses from which the concentrated runoff gradually sinks into the subsurface. In the case of the Pecos River, however, subsurface drainage appears in part to have been pirated away from the stream course by the deeper basins lying nearby and athwart its course.

PRINCIPAL CAUSES OF LOCALIZED EVAPORITE SOLUTION: Tilting of the Delaware Basin to the east, chiefly in late Tertiary time, caused the western part of the evaporite section to be elevated and to some extent exposed, with the consequent removal of the salt and much of the anhydrite, mostly by solution. As solution continued, the western edge of the salt retreated to its present position (Pl. 2), which is considerably east of the western limit of the gypsum and anhydrite.

On the opposite or eastern side of the Delaware Basin, however, where the Capitan reef is buried in the subsurface, solution has been localized in a belt paralleling the reef front. The writers believe that, at the time this

eastern area permanently emerged from beneath epicontinental seas in late Cretaceous time, the beds overlying the Capitan reef tended to have some slight basinward dip, however small. This basinward increment of dip caused downward-percolating groundwaters to move westward toward and over the Capitan reef front. As suggested by Adams (1944, p. 1623), probable slight warping over the Capitan reef tended to open fractures in the overlying bedrock by which waters could gain access to the evaporite section. The exceptionally thick salt deposits encountered in the reef vicinity were readily soluble in the apparently unsaturated waters. With a gradually accelerated rate of solution due to concentration of subsurface drainage, increased amounts of salt were removed, until today there is little or no salt remaining in this narrow belt.

The two separated basins which contain thick fill deposits in the center of the Delaware Basin were initially formed by the local concentration of runoff, possibly influenced by very minor structural undulations, and the downward percolation of the water, which gradually dissolved the underlying salts. It is likely that the nuclei of these two basins were ordinary sinks, which gradually evolved to their present size by incorporating lesser neighbors and pirating other courses of subsurface drainage. Minor basins of interior drainage, of one to possibly several miles in diameter, are known to exist elsewhere in the Delaware Basin; Queen Lake, southeast of Carlsbad, in sec. 30, T. 24 S., R. 29 E., Eddy County, New Mexico, is apparently a small example. Adams (1944, p. 1624) cites Nash Draw, 16 miles southeast of Carlsbad, New Mexico, as a recent and unfilled solution valley that is now developing. This northeast-trending draw has a linear extent of about 18 miles.

For the formation of basins by the dissolution of evaporites, some of the saturated liquids need to be removed. Although the exact methods by which this is accomplished are not entirely clear, some of the liquids are apparently removed by subsurface circulation eventually bringing them near enough to the surface so that they drain into the Pecos River or its subsurface counterparts. Along some stretches of the Pecos River south of Carlsbad in New Mexico, salt crystals have formed on the bed of the stream in and near areas where salty springs are reported.

Structural Features in the Rustler Formation

Any study relating accumulations of fill in the Delaware Basin to evaporite solution would be incomplete without a structural map drawn on the top of the Rustler formation, for the purpose of seeing whether an intermediate horizon reflects the proposed structural relation between the evaporite deposits and the beds of fill. The Rustler (Fig. 2) is the only formation which can be utilized for structural mapping in the interval between the base of the evaporite deposits and the younger surface beds. Its widespread occurrence, distinctive lithologic character, and relatively uniform thickness over most of the area make it an ideal formation for this purpose.

A comparison of the Rustler structural map (Pl. 3) with the fill and salt isopach maps (Pls. 1, 2) makes clear the close relationships between Rustler structure, thickness of salt, and thickness of fill. The small Rustler basins which lie along the Capitan reef front on the eastern side of the Delaware Basin coincide in position with areas of thin or no salt, as well as with areas of thickest fill. A Rustler basin extending northwestward from Loving County, Texas, into New Mexico, marks the position of another area of thin salt and thick fill. A major Rustler basin centered just northwest of the town of Pecos also coincides in position with an area of thick basin fill, beneath which much anhydrite, as well as salt, has obviously been removed by solution.

Extending along most of the eastern side of the Delaware Basin, approximately 12–20 miles west of the Capitan reef front, is a broad anticlinal area of Rustler; the area is slightly arcuate, concave toward the west. This structurally high area of Rustler (Pl. 3) marks a belt of maximum salt thickness (Pl. 2) and minimum fill (Pl. 1). Although the broader relations of Rustler structure to thickness of salt, and also thickness of salt to thickness of fill, are evident, the details in corresponding areas on Plates 1, 2, and 3 may differ. Part of the

difference stems from lack of complete data from individual wells and from the fact that the three maps were contoured independently of each other.

On the southwest side of the Delaware Basin, in Reeves County, a small Rustler basin is present about 8 miles northeast of the Capitan reef front (Pl. 3). There is approximately 500 feet less total evaporites in the Miami Operating Company No. 1 Balmorhea Ranches in the center of this basin than in two adjacent wells about on strike with it, the Wentz No. 1 Dunsaville, 12 miles northwest, and the Ohio No. 1 Popham, 7 miles southeast. Evaporite deposits are also thicker in more basinward wells. These facts seem to indicate that gypsum or anhydrite solution may have influenced the location of this Rustler basin, as no salt is present in the three wells. Thick fill deposits do not appear to be present in this basin, although well control is quite poor.

Thinness or apparent absence of post-Rustler consolidated strata can be noted at several places in the cross sections on Plate 1. This is believed due in part to the intermixing of fill with these beds as they collapsed in the areas where solution was greatest. Hence in some localities it is impossible to separate the formerly consolidated beds from the overlying bodies of basin fill.

Conclusions

This study emphasizes the extreme amount of evaporite solution, particularly of salts, that has occurred in local areas of West Texas and southeastern New Mexico, where thick sections of upper Permian evaporites are abundant. Major structural features in the Rustler formation and major accumulations of fill are directly related to the amount of solution that has taken place in the evaporite beds. In general, in areas of maximum salt thickness, the Rustler formation is structurally high and thickness of fill is at a minimum. Conversely, areas of minimum salt thickness are overlain by structurally low Rustler and the thickest deposits of basin fill. Difficulty in determining the degree to which such structural features in the Rustler formation are due to true tectonic movements or to solution and subsequent slump is obvious. In the opinion of the writers, however, evaporite solution is the most plausible explanation of the conditions described.

References Cited

Adams, J. E. (1936) *Oil pool of open reservoir type*, Am. Assoc. Petrol. Geol., Bull., vol. 20, p. 780–796.

——(1944) *Upper Permian Ochoa Series of Delaware Basin, West Texas and Southeastern New Mexico*, Am. Assoc. Petrol. Geol., Bull., vol. 28, p. 1596–1625.

Horberg, Leland (1949) *Geomorphic history of the Carlsbad Caverns Area, New Mexico*, Jour. Geol., vol. 57, p. 464–476.

Knowles, D. B. and Lang, J. W. (1947) *Preliminary report on the Geology and Ground-Water resources of Reeves County, Texas*, Texas Bd. of Water Engs., 87 pages.

Humble Oil and Refining Co., Box 1271, Corpus Christi, Texas; Box 2180, Houston, Texas.

Manuscript Received by the Secretary of the Society, August 6, 1951.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 64, PP. 547–560, 6 FIGS., 2 PLS. MAY 1953

MUDFLOW OF 1941 AT WRIGHTWOOD, SOUTHERN CALIFORNIA

BY ROBERT P. SHARP AND LAURENCE H. NOBLES

ABSTRACT

During early May of 1941 the southern California mountain resort of Wrightwood was partly inundated by surging flows of muddy debris which occurred daily for more than a week. The immediate cause of flowage was rapid melting of deep winter snow. Intense shattering and rapid weathering of the bedrock (Pelona schist) within the San Andreas rift zone at this location contributed materially to preparing debris for flowage.

Field studies, eyewitness accounts, and movie films provide the following information. Debris was transported 15 miles by the mudflow, and on a gradient as low as 75 feet per mile at the outer extremity. The flow advanced in successive waves or surges of "slimy gray cement-like" mud containing abundant stones. Velocities of the surge fronts ranged from a few to nearly 15 feet per second and averaged close to 10 feet per second during the more fluid phases. The fronts of the more fluid surges slithered and splashed along about like a rapidly advancing tongue of water. No breaker-like motion was evident, although the top of the front tended to shoot ahead of the base. In more viscous surges a bouldery embankment developed at the front and was shoved along by the material following behind.

One sample of the fluid debris had a density of 2.4, indicating a water content of 25–30 per cent by weight. Calculated viscosities range from 2×10^3 to 6×10^3 poises. An average sorting coefficient of 3.94 indicates poor sorting, although somewhat better than in many glacial tills which the flow deposits strongly resemble. Earlier mudflows have occurred here, and others will undoubtedly take place in the future.

CONTENTS

TEXT

	Page
Introduction	547
Introductory statement	547
Location and physical setting	548
Terminology	550
Description of the flow	550
Dimensional relations	550
Data from eyewitness accounts and motion pictures	551
Cause and duration	551
Nature of flowage	551
Velocity of flow	552
Properties of the flow debris	552
Channel modification	553
Mudflow of 1943	553
Source area	553
Nature of mudflow deposits	554
Constitution and characteristics	554
Size relations of constituents	554
Thickness	556
Outer limit of mudflow	556
Attendant features	557
Other mudflows at Wrightwood	558
Summary	558
References cited	559

ILLUSTRATIONS

Figure	Page
1. Map of Wrightwood mudflow and environs	548
2. Map of source area and upper part of Wrightwood mudflow	549
3. Sketch of longitudinal section through advancing mud surge	551
4. Plot of median grain size against distance from source for 1941 mudflow deposits	555
5. Cumulative curves for samples of 1941 mudflow	556
6. Details of mudflow relations	557

Plate	Facing page
1. Slump block and bedrock narrows in source area	552
2. Mudflow terrace and buried cabin	553

INTRODUCTION

Introductory Statement

On May 7, 1941, the resort community of Wrightwood was startled by surging flows of muddy debris which swept down Heath Canyon on the north flank of the San Gabriel Mountains and inundated the outskirts of the village. During the height of activity at midday these waves or surges occurred in rapid succession and consisted of "highly fluid slimy cement-like" mud containing abundant stones.

Some damage resulted, but no persons were injured. The surges occurred daily for more than a week, but the activity ceased at night. The water was supplied through rapid melting of deep snow at the head of Heath Canyon by unseasonably warm weather.

Unfortunately, neither of us was in southern California at the time, but valuable observations and motion pictures were made by engineers and geologists attached to various government agencies dealing with flood-control problems, as well as by local residents and other eyewitnesses who came from considerable distances to look upon this spectacle. Additional information has been obtained from study of the source area, the flow path, and the flow deposits at intervals from 1948 to 1952. An increasing interest in mudflows, their products and geological significance, motivates preparation of this report.

Appreciation is expressed for eyewitness accounts furnished by V. S. Aronovici, Emil Blum, E. L. Hamilton, D. V. Harris, George Richardson, the E. L. Schuylers, M. B. Scott, H. C. Troxell, A. W. Walker, and Royal Ward. Troxell, Aronovici, and Hamilton kindly made available their motion pictures. C. H. Gleason and R. E. Amidon (1941) generously permitted quotation and digest of material in their unpublished report on the Wrightwood flow. Hugo Benioff and H. C. Stetson assisted on special points, and J. P. Buwalda and R. H. Jahns aided with constructive suggestions. All these services are gratefully acknowledged.

Location and Physical Setting

Wrightwood lies 42 airline miles northeast of Los Angeles near the southern edge of Mojave Desert (Fig. 1) in San Bernardino County, California (34° 21.5′ N., 117° 37.5′ W.). The area of the 1941 mudflow is covered by the San Antonio and Shadow Mountains quadrangles. Vertical air photographs taken before and after the flow show part of its path.

In this area the rugged north face of San Gabriel Mountains rises abruptly 4000 to 5000 feet above the gently sloping desert floor. Wrightwood lies within the mountains, south of the northernmost mountain ridge, and near the outlet of Swarthout Valley, an anomalous

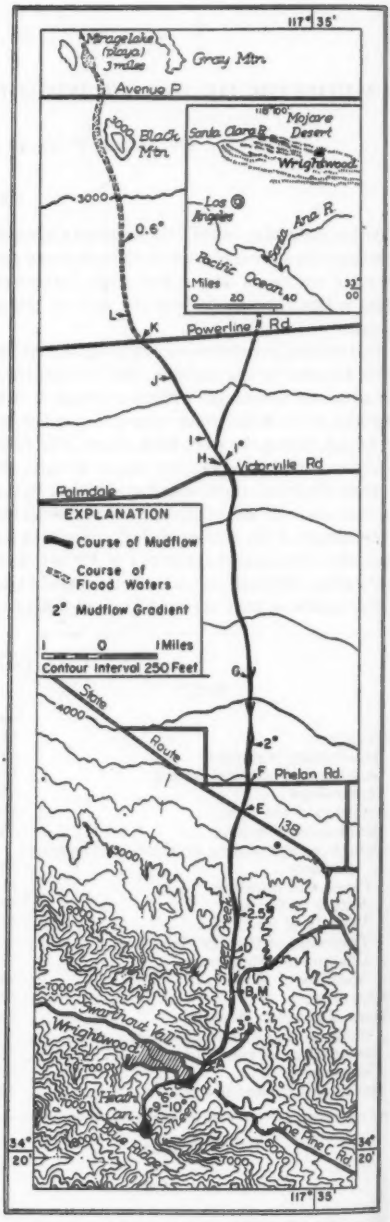


FIGURE 1.—MAP OF WRIGHTWOOD MUDFLOW AND ENVIRONS

longitudinal valley along the San Andreas rift (Noble, 1927; 1932). The eastward continuation of this great fault is marked by the course of Lone Pine Canyon (Fig. 1), and the north face of the mountains is here determined by displacements along subsidiary faults roughly parallel to the San Andreas on the north.

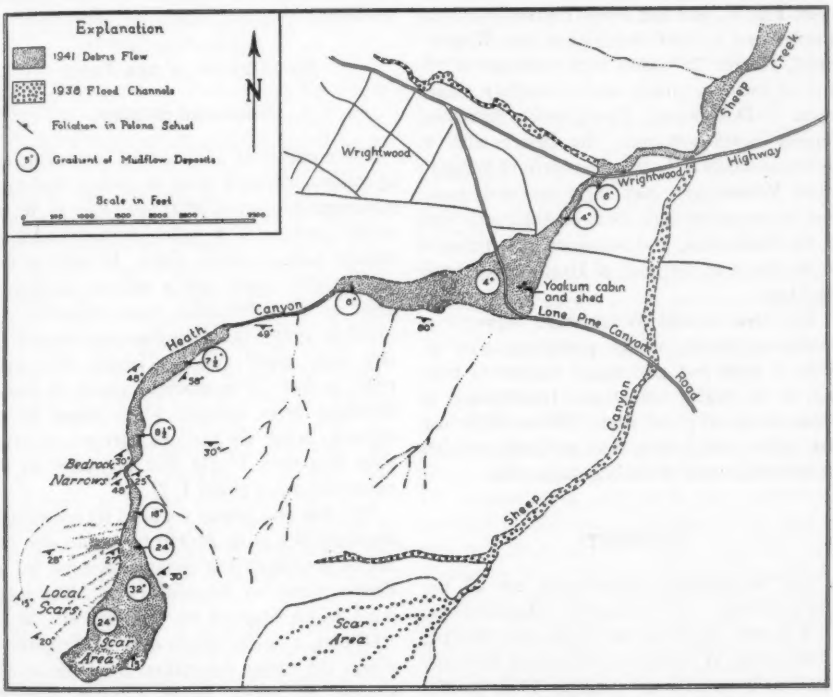


FIGURE 2.—MAP OF SOURCE AREA AND UPPER PART OF WRIGHTWOOD MUDFLOW

The Wrightwood flow originated in a raw scar at the head of Heath Canyon near the southern edge of the San Andreas rift zone. It followed the course of Heath Canyon and subsequently the channel of Sheep Creek. The debris originated at altitudes between 7250 and 8250 feet and descended nearly 5000 feet in reaching its outermost point, 15 miles to the north.

Bedrock in the source area is the Pelona schist (Noble, 1927, p. 28–29) which here consists largely of well-foliated rocks of a variety of textures and compositions. The most abundant rock is a nodular muscovite quartz albite schist. More limited are amphibolite schist, slightly calcareous schist, and a few layers of quartzitic material. Foliation dips consistently into slopes (Fig. 2) from which the debris was derived, but the rock is so badly fractured and sheared, presumably because of proximity to the San Andreas fault, that attitude of the foliation is not particularly significant. Local rumor to the effect that the mudflow was related to an earthquake is not confirmed by records at the Caltech Seismological Laboratory in Pasadena, but comminution of the rock by earlier fault movements is a significant contributing factor.

North of the San Andreas rift are slightly gneissic Mesozoic granitic rocks with inclusions of schist and Paleozoic marble (Noble, 1927, p. 30). Beds of Quaternary fanglomerate lie in fault contact with the older rocks at the edge of the desert, and the desert itself is mantled with younger alluvium swept down from the

mountains. The area north of the rift contributed little debris to the 1941 flow.

The climate of this region is semiarid. Annual precipitation on the mountains just south of Wrightwood is approximately 26 inches (Storey, 1948, Fig. 4), and Big Pines Recreation Area, 4 miles west and 800 feet higher than Wrightwood, records 25 inches with extremes of 15 and 48 inches annually and a monthly maximum of 18.4 inches.[1] Precipitation decreases rapidly northward onto the desert with 8 inches annually only 10 miles north of Wrightwood. Winter snow may linger into early summer in protected high spots on the north side of the mountains, and occasionally it remains in the basin at the head of Heath Creek until mid-July.

The area around Wrightwood supports a coniferous forest, chiefly ponderosa pine at 5700 to 6800 feet and mixed ponderosa pine and fir to higher elevations. Interspersed is some scrub oak, and from 5700 to 4750 feet oak, piñon, and Joshua trees predominate with considerable brush including manzanita.

Terminology

The Wrightwood phenomenon has all the characteristics of a mudflow (Blackwelder, 1928, p. 466), in part of the alpine type (Sharpe, 1938, p. 59). A variety of names has been applied to flows of mud (Sharpe, 1938, p. 57), and the fact that they are not composed solely or even predominantly of mud has given rise to such terms as mud-rock flow (Bailey, Forsling, and Becraft, 1934, p. 1; Marsell, 1949, p. 114–115), mud-rock flood (Croft and Marston, 1950, p. 92), and debris flow (Jahns, 1949, p. 12). Debris flow has also been applied (Gilluly, Waters, and Woodford, 1951, p. 228) to a feature earlier identified as an earth-flow (Blackwelder, 1912, p. 487) and certainly not qualifying as a mudflow.

Debris flow should be used as a general designation for all types of rapid flowage (Sharpe, 1938, p. 49) involving debris of various kinds and conditions. Mudflow is simply a variety of debris flow in which the mud, although not necessarily quantitatively predominant, endows the mass with specific properties and modes of behavior which distinguish it from flows of debris devoid of mud. Thus, the Wrightwood phenomenon is a debris flow of the variety mudflow.

[1] Precipitation records at Big Pines kindly furnished by the Los Angeles County Flood Control District.

Description of the Flow

Dimensional Relations

The mudflow of 1941 carried approximately 15 miles northward from its source, making a maximum descent of 5000 feet (Fig. 1). Water runoff continued 9 miles farther, reaching Mirage Lake, a desert playa. In spite of this considerable length and a volume possibly as great as 1,200,000 cubic yards (Gleason and Amidon, 1941), this flow does not compare in size with great volcanic lahars (Scrivenor, 1929, p. 434). The gradient along its course decreases from roughly 9° in upper Heath Canyon, below the bedrock narrows, to somewhat less than 1° (75 feet per mile) at the outermost point (Figs. 1, 2).

The flow was largely confined by pre-existing channels to a path 20–150 feet wide, and this helped to extend the distance covered, an influence noted by Buwalda (1951). The maximum width mantled by flow deposits, about 1000 feet, is on the debris cone at Wrightwood where the surges debouched from the narrow channel of upper Heath Canyon (Fig. 2). Detritus would have been spread even more widely here except for a flood-control channel and confining embankments constructed after floods of March, 1938. Other embankments hastily thrown up by bulldozers while the 1941 flow was in progress confined it further. Material that crossed the Wrightwood cone was reconcentrated into the narrow channel of Sheep Creek, but it spread out again upon emerging onto the bajada north of the San Gabriel Mountains. Several short lateral lobes and a crude anastomosing pattern were formed below Highway 138, and a width of nearly 300 feet was attained at Phelan road. Farther out on the desert the flow path was narrower, not exceeding 50–150 feet in most places, and with a braided pattern where a single confining channel was lacking. About 0.5 mile below the Palmdale-Victorville road the flow bifurcated,

the main path continuing north-northwest toward Mirage Lake and the side branch extending north-northeast to the vicinity of the Powerline road (Fig. 1).

Data from Eyewitness Accounts and Motion Pictures

Cause and duration.—Actual observation of the flow was limited to the area around Wrightwood and to the channel of Heath Creek up to the source. It is clear from reports of local residents that rapid melting of heavy winter snow during a week of exceptionally warm weather in early May was the immediate cause of flowage. No rain fell just before or during the event. In this respect, the Wrightwood situation resembles that of lofty desert ranges in Asia where numerous mudflows are produced by melting of snow at high altitude (Conway, 1893, p. 292). The flow of Mt. Bandai, Japan, in 1938 was of similar origin (Iida, 1938, p. 681). Lack of flowage during the night at Wrightwood was probably due to reduced nocturnal melting. Surges of muddy debris occurred daily for a week to 10 days from May 7, 1941, and occasional flows may have continued for 2 or 3 weeks. The peak of activity was attained on May 11 and 12.

Nature of flowage.—The debris came down the channel above Wrightwood in a succession of waves or, more appropriately, surges which usually started about 9:00 or 9:30 in the morning, reached a peak of frequency in the early afternoon, and tapered off to an end by late afternoon. Fluidity was greatest at midday when the surges succeeded each other at intervals of a few seconds to tens of minutes. At other times, particularly in late phases of the activity, hours intervened between surges. Gleason and Amidon (1941, p. 3) attribute the surges at Wrightwood to (1) periodic sloughing of debris in the source area, (2) temporary choking of the channel, (3) caving of undercut banks, and (4) friction between the moving debris and the channel. Factors 1 and 2 are considered the most significant.

Successive waves or pulsations are characteristic of mudflows (Blackwelder, 1928, p. 470; Singewald, 1928, p. 482; Jahns, 1949, p. 11–13) and of some water floods (Hovey, 1909b, p. 421). Waves are commonly caused by local choking of the channel with coarse debris and subsequent collapse of the dams as fluid material collects behind them (Conway, 1907, p. 502; Rickmers, 1913, p. 195; Pack, 1923, p. 352; Taylor, 1934, p. 439; Woolley, 1946, p. 80).

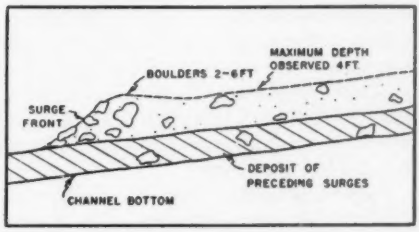


FIGURE 3.—SKETCH OF LONGITUDINAL SECTION THROUGH ADVANCING MUD SURGE
(After Gleason and Amidon, 1941)

E. L. Hamilton's motion pictures clearly illustrate this behavior in the Wrightwood flow and also show that material left in the channel by an earlier surge can be set in motion again. The coming of a new surge was sometimes heralded by the roar of sliding in the source area, and the surge itself made a noise like that of concrete in a mixer. Bumping boulders shook the ground, and the splashing of fluid mud was audible. The bedrock narrows at the foot of the source area was frequently choked by coarse detritus, so that the flowing debris piled up to a thickness of 20–30 feet and spilled over the downstream end of the narrows in a massive cascade, shown in V. S. Aronovici's movie.

In the narrow, confined channel above Wrightwood, the front of an advancing fluid surge at the height of activity slithered and slopped along much like the front of a rapidly flowing tongue of water or the swash from breakers on a beach. There was little evidence of breaker-like rolling under, although the top part tended to override and shoot ahead of the base. The lack of more conspicuous rolling under may have been due in part to "greasing" of the channel by preceding flows. A bouldery embankment formed at the front of more viscous surges (Fig. 3), and the boulders therein rolled, twisted, and shifted about but for the most part did not appear to be rolled under. Instead, they were pushed along by the finer

more fluid debris impounded behind the bouldery dam and swept along by the mud leaking through it. The fronts of fluid surges ranged from a few inches to about 4.5 feet high. The maximum depth measured in the following body of a surge was 4 feet (Gleason and Amidon, 1941, p. 5). As the material became more viscous, particularly in late phases of the activity, velocities decreased, and the bouldery fronts became more massive, steeper, and higher. One such front photographed in motion down a confined channel by H. C. Troxell was between 10 and 15 feet high and consisted largely of boulders 2–3 feet in diameter.

Flow debris accumulated to a maximum thickness of 6 feet over a wide area on the Wrightwood cone near Lone Pine Canyon road, and some surges set up translatory waves less than a foot high which traveled through at least the surficial part of this accumulation. Such waves were observed by D. V. Harris and photographed by V. S. Aronovici.

Velocity of flow.—Surge front velocities measured by competent engineers range from 1 to 2 feet per second to a maximum of 14.5 feet per second. The average of 11 measurements by Gleason and Amidon (1941, p. 4) is 9.4 feet per second, and D. V. Harris (personal communication) reports velocities of 8–11 feet per second. A figure of 9–10 feet per second for the larger surges during the height of activity appears representative. Conway (1893, p. 291) reports speeds of 5 miles per hour (7.4 feet per second) in Himalayan flows.

Properties of the flow debris.—The flowing material was light gray and of mushy to soupy consistency "just like freshly mixed concrete". Much of it consisted of silt, sand, and pebbles less than 1 inch in diameter. The largest boulders were about 6 feet in diameter, and 2- to 3-foot boulders were common near Wrightwood. They were transported in the body of the surges as well as in the snout and clearly did not move in suspension, as sometimes reported "Boulders" bobbing about in the flow proved to be chunks of compact, mud-covered snow from the source area (D. V. Harris and E. L. Hamilton, personal communications).

Bulk density of a sample of fluid debris, determined by San Bernardino County Flood Control engineers, was 2.4, indicating a water content of 25–30 per cent by weight. Maximum solid content of flow debris on the Wrightwood cone was calculated from other sample analyses at 79–85 per cent by weight, indicating a water content of roughly 15–20 per cent. Jahns (1949, p. 12) estimates that the solid material runs as high as 90 per cent in some debris flows. In the laboratory, a sample of the Wrightwood material flowed easily on a 7-degree slope when mixed with approximately 16 per cent water by weight. Density and water content are bound to vary widely, and densities around 2.0 (Woolley, 1946, p. 82) and water contents of 30–60 per cent (Rickmers, 1913, p. 194; Iida, 1938, p. 681) are reported in other flows. Motion pictures of Wrightwood surges at the peak of activity give an impression of high fluidity, even though water content may not have exceeded 30 per cent. Mud was occasionally splashed 20 feet into the air, and spectators on the banks were splattered. At other times the moving material was so viscous that stones thrown onto it caused no splash and sank slowly.

From these data on velocity and bulk density, plus information on thickness and ground slope obtained by field studies, it is possible to make a crude calculation of viscosity in the fluid material at Wrightwood where flowing without the restriction of a confining terminal embankment. The following formula was kindly derived by Dr. Hugo Benioff for this purpose: $n = \frac{\delta g \sin \theta Z_o^2}{2V_s}$ in which n = coefficient of viscosity, δ = density of the fluid debris, g = force of gravity, θ = angle of slope of the ground, Z_o = thickness of the flow, and V_s = velocity at the surface. Using a density of 2.4, a thickness of

PLATE 1.—SLUMP BLOCK AND BEDROCK NARROWS IN SOURCE AREA

FIGURE 1.—TILTED SLUMP BLOCK

Near top and close to east margin of source area at head of Heath Canyon. High-level alluvium exposed in background. Nov. 15, 1948.

FIGURE 2.—DOWNSTREAM END OF BEDROCK NARROWS AT FOOT OF SOURCE AREA

Looking upstream. Material for the mudflow was funneled through this gorge. Letters AA mark mouth of gorge.

FIGURE 1

FIGURE 2

SLUMP BLOCK AND BEDROCK NARROWS IN SOURCE AREA

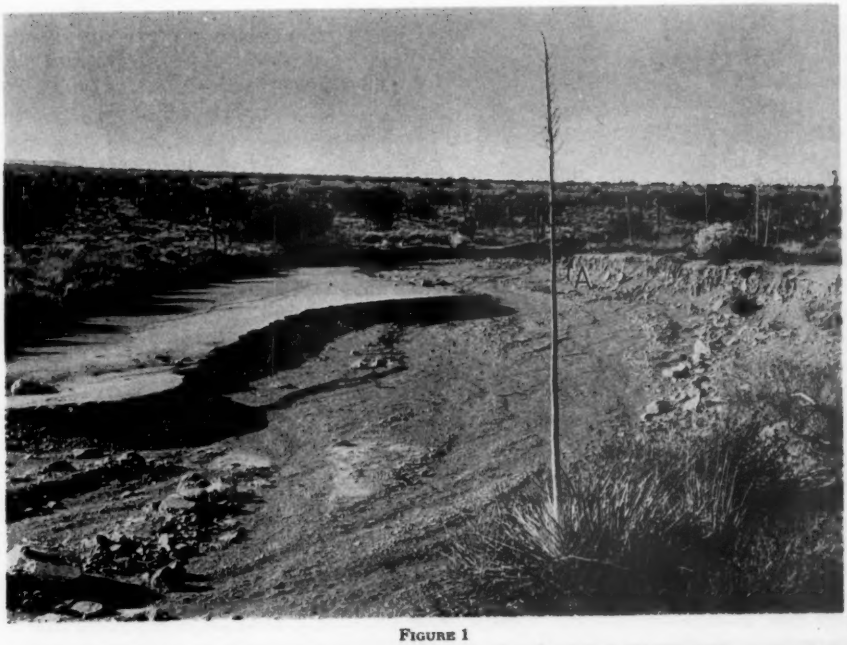


FIGURE 1

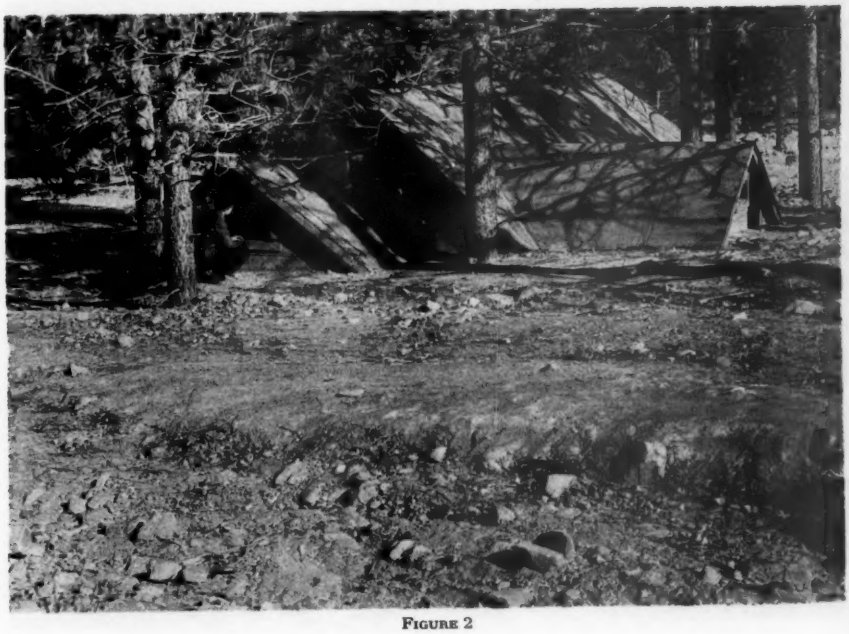


FIGURE 2

MUDFLOW TERRACE AND BURIED CABIN

76.2 cm, a slope angle of 6°, and surface velocities from 300 to 120 cm per second, the coefficient of viscosity ranges from 2.1 × 10^3 to 6 × 10^3 poises. This, of course, is a rough figure as the basic data are crude, and the formula used involves simplifying assumptions among which are Newtonian viscosity, no marginal or terminal influences, no slip on the base and no shear stress on the upper surface, and laminar flow parallel to the base. None of these assumptions need be completely valid, but they are not unreasonable for purposes of approximation, and it is worth nothing that Iida (1938, p. 681) experimentally determined a viscosity of 10^3 poises for flow debris containing 23 per cent water.

Channel modification.—The observations of Gleason and Amidon (1941, p. 6) suggest considerable deepening and some widening of the channel above Wrightwood during the flow, and headward migration of a prominent knickpoint was observed. This erosion must have occurred during the most fluid phases of the activity. Stiff viscous flows occasionally filled the channel to such a degree that overflowage occurred.

Debris flows are often followed and partly dissected by flood water (Conway, 1893, p. 291; Blackwelder, 1928, p. 469–470; Taylor, 1934, p. 440; Troxell and Peterson, 1937, p. 87), but this is not universal (Jahns, 1949, p. 13). No flood waters succeeded the Wrightwood flow, although a modest stream of water ran in the channel for several days after the flow.

Mudflow of 1943.—Moving pictures of debris surges in upper Heath Canyon made almost exactly 2 years later, May 10, 1943, by H. C. Troxell provide further data on flow mechanism and behavior. These surges did not get below the bend of Heath Canyon and were not widely known or observed. They occurred late in the afternoon and came at intervals of about half an hour, each wave being preceded by the noise of sliding in the scar area at the head of Heath Canyon. The surges moved at a speed of several feet a second (2–3 miles per hour), and the channel was flushed out by more fluid material following the stiff viscid snout of each surge. The water, as in 1941, was supplied through melting of heavy winter snow by a spell of warm weather.

Source Area

According to local residents, the great raw scar at the head of Heath Canyon, clearly the source of the 1941 mudflow, was in existence long prior to that event. The geological section shows that it has produced similar flows in times past and will undoubtedly give rise to others in the future. The face of the scar rises 1000 feet at a steep angle, and at its top is a nearly vertical arcuate scarp, 100–150 feet high. Much of the scar is mantled with tough coherent debris, rich in fines, which becomes mobile whenever sufficiently moistened, as shown by small tongues of debris extending across snow banks on the scar area in May, 1949. Shearing and fracturing of the Pelona schist bedrock followed by weathering have produced copious amounts of fine micaceous debris particularly suited for flowage. The scar faces north, and the water supplied by melting snow is supplemented by seeps, emerging from the rock. The abundant moisture facilitates weathering by chemical alteration, and mechanical disintegration is probably promoted by numerous fluctuations across the freezing point.

Small slump blocks, bearing large coniferous trees, dot the scar area, particularly the eastern part. Backward rotation of the blocks is recorded by tilting of the trees at angles up to 28° (Pl. 1, fig. 1). A few trees are tilted forward, presumably because their roots have been loosened so they naturally lean downhill.

Cores were taken from tilted trees on several blocks with the hope that changes in growth related to slumping and tilting would be reflected in the rings, thus dating the movement. This technique has been successfully used on the Gros Ventre slide in Wyoming by Lawrence (1950, p. 246; personal communication). Variations in ring widths recorded at

PLATE 2.—MUDFLOW TERRACE AND BURIED CABIN

FIGURE 1.—MUDFLOW TERRACE ALONG LEFT BANK OF SHEEP CREEK
Just north of Highway 138. Note thin layer of flow debris overlapping right bank at A.

FIGURE 2.—CABIN OF SIDNEY YOAKUM AT WRIGHTWOOD
Filled and buried to the eaves by 1941 mudflow.

Wrightwood are irregular and erratic, giving no consistent picture even for trees on the same block. Probably slumping has reoccurred several different times, and each block may be made up of units which have behaved differently. Possible climatic changes affecting tree growth are a further complication.

At the lower end of the scar is a bedrock narrows 200–300 feet long, 15–20 feet wide at the bottom, and 75 feet deep with extremely steep walls (Pl. 1, fig. 2). All debris from the scar passes through this gorge which functions as a bottleneck. Remnants of debris fills, up to 75 feet thick, upstream from the narrows suggest that material gradually accumulates behind the bottleneck until an exceptional supply of water makes the detritus mobile. V. S. Aronovici's films and observations show that the narrows were periodically choked with coarse detritus during the 1941 flow and that subsequent collapse of the jam under increasing pressure of material accumulating upstream gave rise to a surge of debris travelling rapidly downstream.

Sheep Canyon, east of Heath Canyon (Fig. 2), has a similar headwater scar but without a gorge at its foot. No mudflow came out of upper Sheep Canyon in 1941, probably because sufficient material had not accumulated to generate a flow. The debris appears to be carried away periodically by water floods like that of March, 1938.

Nature of Mudflow Deposits

Constitution and Characteristics

Fines are abundant in the flow deposits, and the numerous large rock fragments are predominantly angular or subangular. Sorting is poor and bedding is nil, except far out on the desert where some crude layering is visible. This structure probably represents a successive superimposition of thin flow units. At Wrightwood, as in other areas, the material looks exactly like some glacial tills (Blackwelder, 1928, p. 465, 469; Bonney, 1902, p. 8). The flow detritus was derived almost wholly from the Pelona schist. Little material was picked up from rocks north of the San Andreas fault, as only occasional fragments of granitic gneiss and marble are seen. Local chunks of oxidized sand and gravel were derived from near-by channel banks through undercutting and caving. Toward its head the flow contains many tree trunks, and elsewhere tin cans and a highway "Danger" sign were incorporated in the deposit.

The Wrightwood mudflow debris can usually be identified by the above characteristics plus a fresh gray color, by topographic forms such as lobate tongues and marginal terraces (Pl. 2, fig. 1), and by abundant bubble holes in the matrix. Bubble holes are not an absolute criterion, for water-laid material may contain bubbles particularly if laid down over dry ground. Nonetheless, at Wrightwood bubbles are more abundant and more uniformly distributed through the fine matrix of the flow deposits than in any of the associated water-laid materials. Abundant inclusions of wood and bark fragments, chips from pine cones, and rabbit droppings also suggest mass flowage, for most of this light material should have been floated away if water were the transporting medium. Rickmers (1913, p. 195) notes that wood fragments, relatively rare in alluvial deposits, are common in the mudflows of Turkestan. Cumulative curves based on mechanical analyses also help distinguish mudflow from water-laid deposits, as shown by comparison of samples K to M with samples A to J (Fig. 5). In some cut banks a distinct unconformity can be seen between the flow debris and underlying alluvium.

Size Relations of Constituents

Larger fragments in the flow debris decrease progressively in size outward from the source, but this is not true of the finer constituents. The largest boulders, 4–6 feet in diameter, are above Wrightwood. At Lone Pine Canyon road the largest stones are 2–3 feet in diameter, and 6–12 inches is the predominant size. At the 5150 contour on Sheep Creek, a lobate debris tongue contains many 12- to 18-inch boulders, and at the Palmdale-Victorville road a similar tongue includes 6- to 12-inch stones. Occasional boulders up to 2 feet in diameter are seen as far out as Phelan road and up to 18 inches at the Palmdale-Victorville road. The 1941

deposit in the main path near the Powerline road is devoid of stones, but concentrations of boulders 4–12 inches in diameter, and occasionally larger to 2 feet, are seen along the north-northeast branch of the flow between the Palmdale-Victorville and Powerline roads. At least some of this coarse material is in an older debris flow, and the stones in the 1941 debris may have been picked up from this earlier deposit. A decrease in both size and amount of coarse debris outward from the source is further demonstrated by convergence of the median grain size values for complete samples and for material of less than 4.76 mm diameter in each sample (Fig. 4).

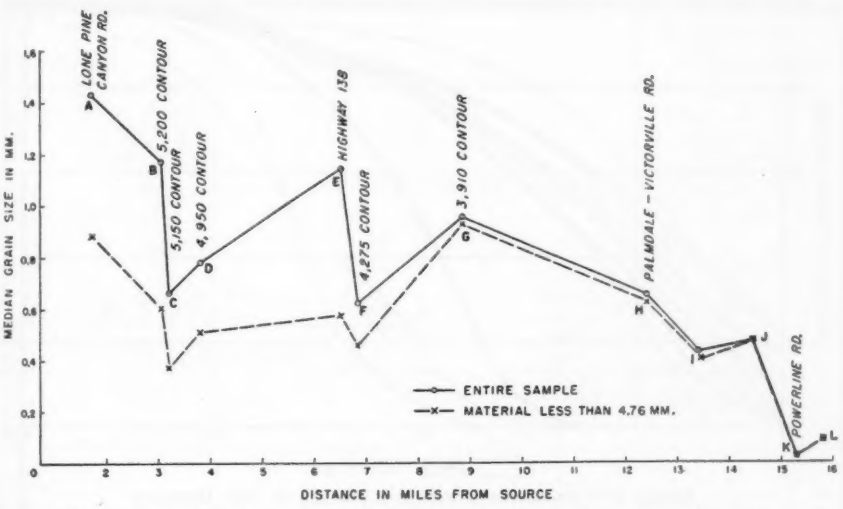


FIGURE 4.—PLOT OF MEDIAN GRAIN SIZE AGAINST DISTANCE FROM SOURCE FOR 1941 MUDFLOW DEPOSITS

The plot of median grain size (Fig. 4) for materials of less than 4.76 mm diameter is noteworthy for its erratic fluctuations and the lack of a consistent decrease outward from the source, at least until the outer limit of mud flowage is passed. This is logical, for a flow has no obvious means of sorting its fine constituents, only the larger stones are gradually eliminated, giving a decrease in median size for the whole sample but not for the finer material. The erratic fluctuations in median grain size simply reflect the longitudinal inhomogeneity of the flow deposit, composed as it is of many individual debris tongues which differ from each other and which are not homogeneous within themselves.

Frequency curves for the 1941 mudflow debris are nearly all bimodal with the second and smaller maximum in the neighborhood of 0.025 mm. The frequency curve for a Utah flow (Crawford and Thackwell, 1931, p. 102–103) also displays a secondary maximum at about 0.037 mm, and Chawner (1935, p. 261) obtained a bimodal diagram from possible debris-flow deposits of the Montrose flood. Most bimodal curves are attributed to multiple sources of debris, and polymodal curves obtained in poorly sorted till presumably resulted from mixtures of debris from several places (Krumbein, 1933, p. 389–390; Deane, 1950, p. 24). At Wrightwood, multiple sources are not indicated by other relations, and the cause of the bimodal characteristic remains a mystery. It may be a function of grain-size distribution in the source rock or a product of peculiarities in disintegration of that rock. Actually the second maximum is so small that it would not attract attention were it less consistent.

Mechanical analyses confirm other relations

more or less evident from field study. The sorting ratio (S_o) is high, ranging from 2.67 to 5.03 with an average of 3.94. This is better than the sorting in some tills (Krumbein, 1933, p. 391; Krumbein and Pettijohn, 1938, p. 231–232; Deane, 1950, p. 16–17) with an average ratio of about 5. Considerable differences must be expected between and within various flow deposits, and an analysis of flow debris in Utah (Crawford and Thackwell, 1931, p. 102–103) yields a cumulative curve with a higher degree of sorting, therefore a lower sorting ratio (S_o = 2.76), than at Wrightwood. The source of the Utah debris, a sandy terrace of Lake Bonneville, is clearly a significant factor.

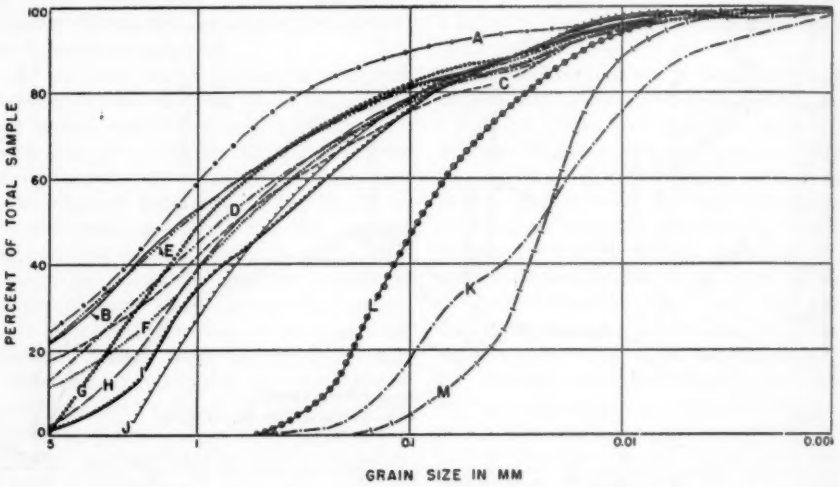


FIGURE 5.—CUMULATIVE CURVES FOR SAMPLES OF 1941 MUDFLOW
For locations of samples see Figure 1. Samples A to J are mudflow materials, and samples K to M are waterlaid detritus.

Thickness

The thickness of flow deposits decreases outward from the source but with great variation owing to accumulation in irregularities of the ground or behind obstructions. In the source area, above the bedrock narrows, remnants of debris fills indicate a thickness of 50–75 feet, and chunks of flow debris cling to the walls of the narrows 15–20 feet above its floor. Downstream the thickness decreases rapidly, attaining a maximum of 6 feet in the vicinity of Wrightwood and not more than 2–4 feet along the course of Sheep Creek. The maximum thickness at Phelan road is about 4 feet, at the Palmdale-Victorville road it is 20 inches and somewhat less at the Powerline road. Older debris-flow deposits are exposed in places beneath the 1941 debris (Fig. 6, II) and give an erroneous impression of thickness unless properly identified. This proved especially troublesome at the Palmdale-Victorville road where the 1941 debris was first thought to be 3 feet thick. Subsequent discovery of an intercalated layer of oxidized wind-blown silty sand showed that this thickness included both the 1941 and older flow deposits. Other exposures of the older flow debris were subsequently found between the Palmdale-Victorville and Powerline roads.

Outer Limit of Mudflow

Determination of the outermost limit of mud flowage, the point at which mass movement gave way to water transport, is difficult, but the cumulative curves from mechanical analyses give considerable aid and confirm conclusions based on field observations. Along the main path, the change occurred about 0.5 mile above

(south of) the Powerline road. The cumulative curves (Fig. 5) for samples H, I, and J between the Palmdale-Victorville and Powerline roads strongly resemble those of unquestionable flow debris. By comparison, the curves for samples K and L near the Powerline road are strikingly similar to that of a water-laid silt (sample M, Fig. 5). Cumulative curves for samples collected along the north-northeast branch indicate mass movement almost to the Powerline road. Thus, flowage occurred along both the main path and the north-northeast branch for approximately 15 miles from the source at the head of Heath Canyon.

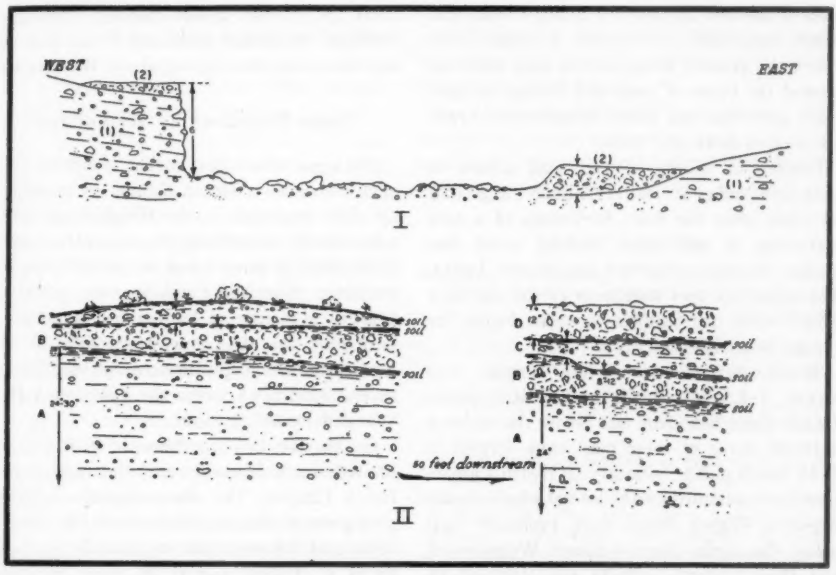


FIGURE 6.—DETAILS OF MUDFLOW RELATIONS

I. Overtopping of high bank on outside of curve
(1)-Alluvial gravel
(2)-1941 mudflow deposits
II. Relation with alluvial gravels and older mudflow deposits
A—Oxidized alluvial gravel with prominent soil layer at top
B—Pre-1941 mudflow deposit with dark soil layer at top
C—Local alluvial deposits derived from nearby slopes with weak soil layer and brush at top
D—1941 mudflow deposits, fresh, gray, no soil and burying brush on top of alluvial material (C)

ATTENDANT FEATURES

Some damage resulted to highways, roads, buildings, and other works of man. At least 3 cabins and several garages or sheds at Wrightwood were damaged, two of the cabins beyond repair. A cabin and shed, owned by Sidney Yoakum, about 50 yards below the Lone Pine Canyon road were buried to the eaves by debris (Pl. 2, fig. 2). Curiously, neither building appears to have been moved nor to have suffered damage other than burial and filling, probably because of low impact velocity and homogeneous flow. Similar occurrences elsewhere are reported (Troxell and Peterson, 1937, p. 87). The walls of other buildings at Wrightwood were said to be partly caved in. Warnings were ample so that no one was trapped, and most furniture was removed from the cabins. A small fruit orchard at Wrightwood was ruined by the flow, many trees being broken down or killed by partial burial. Roads were buried, but paving for the most part was not seriously damaged, a performance rather typical of mudflows (Blackwelder, 1928, p. 468).

In upper Heath Canyon many large trees were uprooted, broken down, or so deeply buried at the base that they subsequently died. Many instances of severe battering of tree trunks can still be seen 2–3 feet above the level of the flow deposits. A 5-inch cobble was deeply embedded in the trunk of a pine 4 feet above the ground. Large stones were piled up around the bases of trees and bushes to form crude garlands, and below Wrightwood, brush was beaten down and buried.

Patches of splattered mud still adhere to trees and bushes several feet above the ground 10 years after the flow. Remnants of a thin plastering of gray mud, looking much like gunite, are also preserved on channel banks. The coherence and toughness of the fine flow debris when dry justifies to some degree the comparison with cement.

Boulder-rich, lobate debris tongues and levees, 1–1.5 feet high, are evident where Heath Canyon widens out below the bedrock narrows. Some of these may have formed in 1943. Small gullies tributary to Heath Canyon have been dammed by the lateral edges of such deposits. Higher levees were probably built along the main channel above Wrightwood, but these natural features are obscured by embankments built by bulldozers.

Striations and scratches on stones and bedrock produced by mud flowage were searched for assiduously but without much reward. Bedrock exposures in the narrows and a few boulders farther down the channel have some suggestive marking, but they are not diagnostic and do not compare with striations reported in other flow deposits (Hovey, 1909a, p. 413; Pack, 1923, p. 353; Scrivenor, 1929, p. 434; Blackwelder, 1930).

Small terraces have been formed by dissection of the flow deposits where the pre-existing channel was wide, flat-floored, and not filled to overflowing by debris (Pl. 2, fig. 1). The terrace surfaces are remarkably smooth, being broken only by minor surficial markings and the tops of partly buried boulders. The channel cut into the terraces is floored with gravel and boulders, the fines having been carried away by running water.

In going around a curve, the surges of debris piled up against the outside bank. In many places a bank 5–6 feet high was overtopped and a sheet of fine material 6–12 inches thick spread beyond, while the surface of the flow on the inside of the curve was at a much lower level (Fig. 6, I). Troxell and Peterson (1937, p. 85–86) report similar "toboggan sledding" on a larger scale, and it was observed in action along the channel above Wrightwood.

Other Mudflows at Wrightwood

The expectation that earlier mudflows must have come down Heath Canyon is confirmed by older materials in the Wrightwood debris cone closely resembling the deposits of 1941. In the bank of Sheep Creek at the 4950 contour, remnants of an older mudflow are preserved beneath a mantle of local alluvium which is buried by the 1941 flow debris (Fig. 6, II). Earlier mudflow deposits were also identified at several localities between the Palmdale-Victorville and Powerline roads.

On May 10, 1943, mudflows of small size and limited extent descended into the upper part of Heath Canyon. The climatological conditions giving rise to this occurrence were like those of 1941, and the event was successfully predicted by H. C. Troxell and M. B. Scott. This flow did not get below the bend of Heath Canyon above Wrightwood.

There is every reason to expect that subsequent mudflows will descend from the head of Heath Canyon whenever proper conditions arise, specifically: heavy winter snow and a generally cool spring followed by an exceptionally warm period of several days duration. These flows will probably not be as large as those of 1941 unless debris once again accumulates in large volume upstream from the bedrock narrows at the base of the scar area.

Summary

The significant points derived from study of the 1941 mudflow at Wrightwood are as follows:

1. Material was transported approximately 15 miles by mass flowage.

2. The gradient at the outer end of the flow is less than 1°, about 75 feet per mile, compared to slopes of 24°–32° in the source area and 9° in upper Heath Canyon.

3. The mudflow deposits are an unbedded mixture of poorly sorted, fine to coarse debris incorporating battered fragments of wood, bark, pine cones, and other light materials which should have floated away in running water. The fine matrix contains many entrapped air bubbles giving a pseudo-vesicular structure.

4. Sizes of larger stones in the debris decrease irregularly outward from the source, but the median size of constituents less than 4.76 mm shows no regular decrease with distance. This is consistent with the mechanics of mass flow.

5. The average sorting coefficient (S_o) of 3.94 indicates a poor degree of sorting although somewhat better than in many glacial tills. Cumulative curves help to distinguish mudflow deposits from water-laid materials.

6. From eyewitness accounts, the flow advanced as a series of muddy debris waves or surges succeeding each other at intervals of a few seconds to several hours. Activity attained a peak in the early afternoon and appears to have ceased at night, probably because of decreased melting of snow.

7. At the height of activity, the larger more fluid surges traveled 9–10 feet per second, some attained 14.5 feet per second, but many were slower and velocities dropped to 1–3 feet per second as activity decreased.

8. The front of an advancing fluid surge was like the front of a rapidly flowing tongue of water, but, with increased viscosity, a terminal bouldery embankment developed and was shoved along by the more fluid material behind.

9. A sample of the fluid debris had a density of 2.4, indicating a water content of 25–30 per cent by weight.

10. Viscosity of the flow is calculated at roughly 2×10^9 to 6×10^9 poises.

11. A house and shed near the margin of the flow at Wrightwood were buried to the eaves and filled to approximately the same level inside without other apparent damage, and several other buildings were partly buried.

12. The various situations favorable to development of mudflows have been summarized by Rickmers (1913, p. 195), Blackwelder (1928, p. 478–479), and Sharpe (1938, p. 56). Here the essential conditions are: a raw slide area at the head of Heath Canyon, a badly sheared and shattered bedrock which yields much fine and poorly sorted micaceous debris upon weathering, a narrow rock gorge at the base of the source area behind which debris accumulates, and a north-facing slope upon which snow gathers deeply during winter and remains into late spring and early summer when it may, at times, be melted rapidly by an unseasonable warm spell.

13. Similar mudflows have occurred at this locality before and since 1941, and it can be predicted with confidence that others will come down from upper Heath Canyon in the future.

References Cited

Bailey, R. W.; Forsling, C. L.; and Becraft, R. J. (1934) *Floods and accelerated erosion in northern Utah*, U. S. Dept. Agri., Misc. Pub. 196, 21 pages.

Blackwelder, Eliot (1912) *The Gros Ventre slide, an active earth-flow.* Geol. Soc. Am., Bull., vol. 23, p. 487–492.

——(1928) *Mudflow as a geologic agent in semi-arid mountains*, Geol. Soc. Am., Bull., vol. 39, p. 465–480.

——(1930) *Striated boulders as evidence of glacial action*, Jour. Geol., vol. 41, p. 154.

Bonney, T. G. (1902) *Moraines and mud-streams in the Alps*, Geol. Mag., vol. 9, p. 8–16.

Buwalda, J. P. (1951) *Transportation of coarse material on alluvial fans*, (Abstract) Geol. Soc. Am., Bull., vol. 62, p. 1497.

Chawner, W. D. (1935) *Alluvial fan flooding, the Montrose, California, flood of 1934*, Geog. Rev., vol. 25, p. 255–263.

Conway, W. M. (1893) *Exploration in the Mustagh Mountains*, Geog. Jour., vol. 2, p. 289–299.

——(1907) *Comment on "The Fan Mountains in the Duab of Turkestan by W. R. Rickmers"*, Geog. Jour., vol. 30, p. 501–502.

Crawford, A. L.; and Thackwell, F. E. (1931) *Some aspects of the mudflows north of Salt Lake City, Utah*, Utah Acad. Sci., Pr., vol. 8, p. 97–105.

Croft, A. B.; and Marston, R. B. (1950) *Summer rainfall characteristics in northern Utah*, Am. Geophys. Union, Tr., vol. 31, p. 83–93.

Deane, R. E. (1950) *Pleistocene geology of the Lake Simcoe district, Ontario*, Geol. Survey Canada, Mem. 256, 108 pages.

Gilluly, James; Waters, A. C.; and Woodford, A. O. (1951) *Principles of geology*, W. H. Freeman and Co., San Francisco, 631 pages.

Gleason, C. H.; and Amidon, R. E. (1941) *Landslide and mudflow, Wrightwood, California*, Calif. Forest and Range Experiment Sta., Unpub. rept., p. 1–7.

Hovey, E. O. (1909a) *Striations and U-shaped valleys produced by other than glacial action*, Geol. Soc. Am., Bull., vol. 20, p. 409–416.

——(1909b) *Clearing out of the Wallibu and Rabaka gorges on Saint Vincent Island*, Geol. Soc. Am., Bull., vol. 20, p. 417–426.

Iida, Kumizi (1938) *The mud flow that occurred*

near the explosion crater of Mt. Bandai on May 9 and 15, 1938, and some physical properties of volcanic mud, Tokyo Imp. Univ., Earthquake Res. Inst., Bull. 16, p. 658–681 (Jap., Engl. Summ.).

Jahns, R. H. (1949) *Desert floods*, Eng. and Sci. Jour., vol. 12, p. 10–14.

Krumbein, W. C. (1933) *Textural and lithological variations in glacial till*, Jour. Geol., vol. 41, p. 382–408.

——; and Pettijohn, F. J. (1938) *Manual of sedimentary petrography*, New York, D. Appleton-Century Co., 549 pages.

Lawrence, D. B. (1950) *Estimating dates of recent glacier advances and recession rates by studying tree growth layers*, Am. Geophys. Union, Tr., vol. 31, p. 243–248.

Marsell, R. E. (1949) *The Quaternary system in Utah*, in *The oil and gas possibilities of Utah*, Utah Geol. and Min. Survey, Spec. Bull., p. 109–118.

Noble, L. F. (1927) *The San Andreas rift and some other active faults in the desert region of southeastern California*, Seismol. Soc. Am., Bull., vol. 17, p. 25–39.

——(1932) *The San Andreas rift in the desert region of southeastern California*, Carnegie Inst. Year Book, vol. 31, p. 355–363.

Pack, F. J. (1923) *Torrential potential of desert waters*, Pan Am. Geol., vol. 40, p. 349–356.

Rickmers, W. R. (1913) *The Duab of Turkestan*, Cambridge Univ. Press, 563 pages.

Scrivenor, J. B. (1929) *The mudstreams ("Lahars") of Gunong Keloet in Java*, Geol. Mag., vol. 66, p. 433–434.

Sharpe, C. F. S. (1938) *Landslides and related phenomena*, Columbia Univ. Press, 137 pages.

Singewald, J. T. (1928) *Discussion of "Mudflow as a geologic agent in semiarid mountains by Eliot Blackwelder,"* Geol. Soc. Am., Bull., vol. 39, p. 480–483.

Storey, H. C. (1948) *Geology of the San Gabriel Mountains, California, and its relation to water distribution*, Calif. Div. Forestry, 19 pages.

Taylor, C. A. (1934) *Debris flow from canyons in Los Angeles County flood*, Eng. News-Record, vol. 112, p. 439–440.

Troxell, H. C.; and Peterson, J. Q. (1937) *Flood La Cañada Valley, California*, U. S. Ge Survey, W. S. P. 796-C, p. 53–98.

Woolley, R. R. (1946) *Cloudburst floods in Uta 1850–1938*, U. S. Geol. Survey, W. S. Pap 994, 128 pages.

California Institute of Technology, Pas dena, Calif.; Northwestern Universit Evanston, Illinois

Manuscript Received by the Secretary the Society, July 7, 1952

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 64, PP. 561-580, 8 FIGS., 3 PLS. MAY 1953

SYSTEMATIC DRAINAGE CHANGES IN THE PIEDMONT OF NORTH CAROLINA AND VIRGINIA

By William A. White

Abstract

In the Piedmont of North Carolina and Virginia, northeast-flowing tributaries of southeast-flowing rivers such as the Catawba, Yadkin, and James are long and subparallel. Opposing tributaries on the northeast sides of the main streams are short and irregular in direction. This asymmetry suggests the former existence of a subsequent drainage flowing northeastward toward the Potomac. Apparently it was dismembered by multiple captures effected by short southeast-flowing streams of steeper gradient.

The drainage area of this former stream system is delimited by an enclosing zone of monadnocks on the southeast and southwest, and by the Blue Ridge Scarp on the northwest. It is developed largely upon the crystalline rocks, while the marginal monadnocks (Uwharie Hills, etc.) which mark the former divide on the southeast are developed largely on the Carolina slates. The zone of terminal monadnocks which marks the headwater divide extends across the strike, on the crystalline rocks approximately along the boundary between North Carolina and South Carolina.

Apparently, a succession of captures progressed across the area as the stronger southeast-flowing streams insinuated their headwaters northwestward, dismembering and diverting each of the northeast-flowing streams in turn, beginning at the southeast and ending at the foot of the Blue Ridge on the northwest. Elbows of capture are sharper in the northwestern streams where captures were later. There, also, gravel-coated strath terraces are preserved across the divides between present major streams such as the Yadkin and Catawba.

Present stream profiles show lower gradients in the area of the former northeast-flowing drainage than in the slate belt to the east.

CONTENTS

TEXT

	Page
Acknowledgments	562
Triassic influence on drainage	562
Asymmetry of drainage	563
Dismemberment of former drainage	563
Strath terraces	564
General location	564
Yadkin-Catawba-Dan terrace	564
Roanoke-James terrace	568
Rockfish-Mechum-Rapidan terrace	568
Appomattox-James terrace	568
Description of gravels	569
Topographic expression of former drainageways	570
Peripheral monadnocks	571
Profiles along present streams	572
Relation to Coastal Plain	574
Other possibilities for establishment of northeast-flowing headwater reaches	574
Fault-angle valleys	574
Differential headward erosion	575
Superposition from alluvium	575
Exhumation of old valleys from beneath Coastal Plain	575
Superposition from coast-parallel consequents	576
Control by plunging structures	576
Relation to drainage north of the Potomac	576
References cited	579

ILLUSTRATIONS

Figure	Page
1. Panoramic view (180°) of valley of Yadkin River from point opposite confluence with Buffalo Cove Creek on southeast side of valley	565
2. Profile along old strath terrace across present drainage areas of Catawba, Yadkin, and Dan rivers	566
3. Profile along old strath terrace across present drainage areas of Snow Creek, Pigg, Roanoke, and James rivers	567
4. Profile along old strath terrace across present drainage areas of Rockfish, Mechum, Rivanna, and Rapidan rivers	569
5. Diagrammatic sketch showing the genesis of eastward slope of Piedmont surface	571

Figure	Page
6. Profiles of present streams across Piedmont of North Carolina	573
7. Configuration of surface of basement rocks beneath Coastal Plain	577
8. Isopach map of "Raritan"-Potomac group	578

Plate	Facing page
1. Drainage map of parts of North Carolina and Virginia	561
2. Strath terraces	570
3. Topographic model and gravel bed	571

Acknowledgments

A large share of the expense incurred in this study was defrayed by a grant from the Carnegie Research Fund and the writer owes much to Dr. E. E. Blodgett for his kind assistance in preparing the manuscript. He is also indebted to Dr. C. S. Maurice and Messrs. V. R. Geyer, R. E. Hall, Linn Hoover, H. M. Peek, M. C. Powers, J. P. Spillers, and R. T. Wolf, who assisted in field work and construction of profiles.

Triassic Influence on Drainage

One of the classic problems of Appalachian geomorphology is the assumed reversal of drainage from the northwest-flowing streams which drained Appalachia to the southeast-flowing streams which drain most of the present Appalachian Mountains to the Atlantic Ocean. The writer has always considered this a problem of small moment because the earlier drainage of the region was probably supplanted by centripetal drainage during the long periods of dry climate which seem to have prevailed during Permian, Triassic, and possibly Jurassic time.

During the period of centripetal drainage, each fault trough was probably broken up into a number of small basins with low divides between them. Return to humid conditions would permit the expansion of playa lakes until these divides were overtopped and a continuous stream developed along the axis of the fault trough. This newly integrated stream would most plausibly escape to the east, away from the newly uplifted Appalachians to the near-by Atlantic.

Triassic structural features, common throughout the Atlantic Seaboard, are indicated by the several remnants of sediment-filled fault troughs and high-angle faults in the intervening basement rocks. Most of these structures (south of the Hudson River) are oriented roughly parallel with the coastline, and perpendicular to the present courses of the arterial rivers, such as the Delaware, Susquehanna, Potomac, James, Roanoke, Peedee (Yadkin) and Santee.

There is ample evidence that these structures had extensive surficial manifestation at the time of their activity, and the extensive fanglomerates along Triassic border faults leave little doubt that such structures dominated their contemporary topography. It is therefore plausible that they left a heavy impress upon drainage patterns which should be difficult to erase entirely.

Carlston (1946) has shown that certain of the present major transverse streams did not exist in Triassic time, and it would therefore seem probable, barring regional superposition from Cretaceous sediments (Johnson, 1931), that the present drainage system has an unbroken line of descent from the streams which began as longitudinal consequents in the Triassic fault troughs.

The long interval between the Triassic and the present should have modified such a drainage, but its ancestral influence might not be wholly obliterated as long as the region preserved a subaerial situation, a climate sufficiently humid to prevent a return to centripetal drainage, and enough fall in the streams to prevent their inordinate lateral wandering.

Not only would the original topographic depressions exercise a strong influence on stream orientation which could be long preserved in its general trend, but, even after several cycles of erosion, the soft sediments of the Newark Series and the structural weakness afforded by the multiple faulting should exert a strong influence in the same direction, tending to preserve a later subsequent drainage in the same avenues as the original Triassic drainage which was consequent to the fault troughs.

Because of the long time since the Triassic period when they were conceived, one would

expect to find only remnants of these ancient drainageways preserved as components of the present drainage. Uplift along the Appalachian axis with southeastward tilting of the Atlantic slope would have relegated any strike streams to positions of secondary importance in a system dominated by headward-extended consequents which flow directly to the sea in courses normal to the coastline.

What appear to the writer to be remnants of such a strike drainage may be seen in the Piedmont of North Carolina and Virginia. A large share of the total stream length of this area is oriented roughly parallel with the coast line and the regional trend of the rock structures (Pl. 1).

Asymmetry of Drainage

There would be little of note about such strike streams if they were approximately equal in length in opposite directions, for subsequence would produce such a pattern. However, the strike streams are prominently asymmetric in relation to their transverse trunks. In the Piedmont of North Carolina and Virginia, the northeast-flowing tributaries of the transverse trunk streams are generally longer and more prominently subparallel than are the opposing, southwest-flowing tributaries on the opposite sides of the trunk streams (Pl. 1). Repeatedly the heads of the northeast-flowing tributaries of one stream system are found within a few miles of the transverse trunk of the adjacent system to the southwest. This relationship can be noted readily between the James and the Roanoke; the Roanoke and the Yadkin; the Yadkin and the Catawba.

Particular attention should be paid to the following northeast-flowing streams, all of which are much longer than the opposing southwest-flowing tributaries on the northeast sides of the transverse trunk streams: (1) The Catawba River above the bend near Catawba, North Carolina, (2) the Yadkin River above Donnaha, (3) Troublesome, Haw, Reedy Fork, Buffalo, and Alamance creeks of the Haw River, (4) the Dan River and its tributaries, Wolf Island, Hogan, Moon, Country Line, Hyco, Sugar Tree, and Aarons creeks, (5) the Banister River above Elkhorn Creek and its tributaries, Elkhorn and Sandy creeks, (6) Snow Creek-Blackwater River and Pigg River tributaries of the Roanoke River, (7) the Appomattox River above Sunnyside and all of its tributaries from the south side (there are none of any consequence on the north side), and (8) the James River above Scottsville, or its confluence with the Hardware River and its tributaries below Scottsville; Slate River, Willis River, and Deep Creek.

Dismemberment of Former Drainage

Such consistent asymmetry suggests the dismemberment of a former stream system which flowed northeastward parallel with the principal geologic structures, and at right angles to the present transverse trunk streams.

It seems plausible that the dismemberment took place through multiple captures by southeast-flowing, headward-extended consequents, which had the advantage of courses perpendicular to the coastline. Such courses would have been shorter and steeper than those of the strike streams, which appear to have flowed long distances parallel with the coastline before reaching it.

Such a series of captures would have begun at the southeastern edge of the former northeast-flowing drainage and progressed systematically across it towards its northwestern edge. Thus the captures would be progressively younger as one crossed the old drainage basin from the southeast to the northwest. It is therefore of interest that the elbows of capture are commonly more rounded near the southeastern edge of the area, and more sharply angular toward the northwestern edge. This variation may well be due to factors other than age, for the area has more rugged topography near its western edge and more prominent structural trends. The classic idea of the elbow of capture as being a right angle stems from two common situations which frequently produce such right angles: (1) the capture of streams which flow parallel with the strike in steeply dipping sediments, or (2) the capture of streams which flow along the back slopes of cuestas close to and parallel with the crests of their scarps. In both these situations the capturing stream is usually insinuated at right angles to the dismembered stream.

On the other hand, in a region of relatively homogeneous rocks, or where there is no prominent parallelism of structural trends, there is little compulsion that the capturing stream approach the dismembered stream at a right angle. In much of the Piedmont the elbows of capture should tend to be greater than 90° because the regional slope which gives the capturing stream its advantage is about east, and the dismembered streams flowed northeastward before capture.

Strath Terraces

General Location

A more considerable argument for the dismemberment of a former drainage system is found in the strath terraces which can be traced with unbroken continuity across the divides between present stream systems. Most of these are fairly close to the northwestern margin of the former northeast-flowing drainage area, close to the foot of the Blue Ridge Scarp.

Yadkin-Catawba-Dan Terrace

Perhaps the most clearly marked of these strath terraces is found in the valleys of the Yadkin, Catawba, and Dan drainages (Fig. 2). Preserved extensively throughout the Morganton and Wilkesboro basins of western North Carolina, it is conspicuous in the valley of the South Fork of the Catawba River where many small steep-sided hills have summits at the same level (Pl. 2, fig. 1). It can be traced from Old Fort at the head of the present floodplain on the South Fork of the Catawba River downstream to the confluence with the North Fork, and thence in the same northeasterly direction up such southwest-flowing tributaries of the Catawba as Lower Creek and Mulberry River.

Lower Creek flows past the city of Lenoir, North Carolina, where the old strath is conspicuous (Pl. 2, fig. 2). Upstream from Lenoir, to the northeast, this same terrace can be traced continuously across the headwater divide where it passes unbroken into the valley of Kings Creek, a northeast-flowing tributary of the Yadkin River. The old surface is continuous throughout the length of Kings Creek to its confluence with the Yadkin, and persists to the northeast down the valley of the Yadkin itself (Fig. 1).

The same surface can also be traced unbroken across the Catawba-Yadkin divide in a strath which persists up Mulberry River of the Catawba system, across the divide at Happy Valley and into the valley of the Yadkin River (The extreme left edge of Figure 1 shows this strath terrace in the upper Yadkin valley.)

All these strath remnants show old channel gravels preserved upon their surfaces. Such gravels can be found on the headwater divides mentioned above, between streams of the two separate drainage systems of the Catawba and Yadkin Rivers, suggesting that the valleys now on opposite sides of these divides once belonged to the same stream.

There is good evidence that this former stream system flowed northeastward. The strath terrace at the head of the present floodplain in the South Fork of the Catawba River ends at Old Fort, North Carolina, in a narrow mountain valley surrounded at its head by high, unborken mountain walls with no opening whereby the old drainage could have escaped to the southwest. On the other hand, the level surface of the old strath terrace persists northeastward from Old Fort all the way across the Catawba drainage and across the present divides where the headwaters of the Catawba tributaries oppose those of the Yadkin.

This same strath terrace persists farther northeast down the Yadkin River with increasing width as the headwater valley widens out into the broad intermontane of the Wilkesboro Basin to become the general surface of the Piedmont Peneplain. On this broad regional surface, the confines of former drainageways are less clearly marked because they lack the confining walls of the mountains to delimit the former floodplains. Therefore, it is more difficult to show that the old drainage continued still further northeast beyond the point where the Yadkin now turns southeastward at Donnaha, North Carolina. However (Fig. 2; Pl. 1) this reach of the Yadkin is closely in line with the Dan River and its tributary, Town Creek, a short distance to the northeast, where they flow northeastward along the axis of the Dan

FIGURE 1.—PANORAMIC VIEW (180°) OF VALLEY OF YADKIN RIVER FROM POINT OPPOSITE CONFLUENCE WITH BUFFALO COVE CREEK ON SOUTHEAST SIDE OF VALLEY

Shows present flood plain, 1500-foot strath terrace and foot of Blue Ridge Scarp. Catawba-Yadkin divide at extreme left.

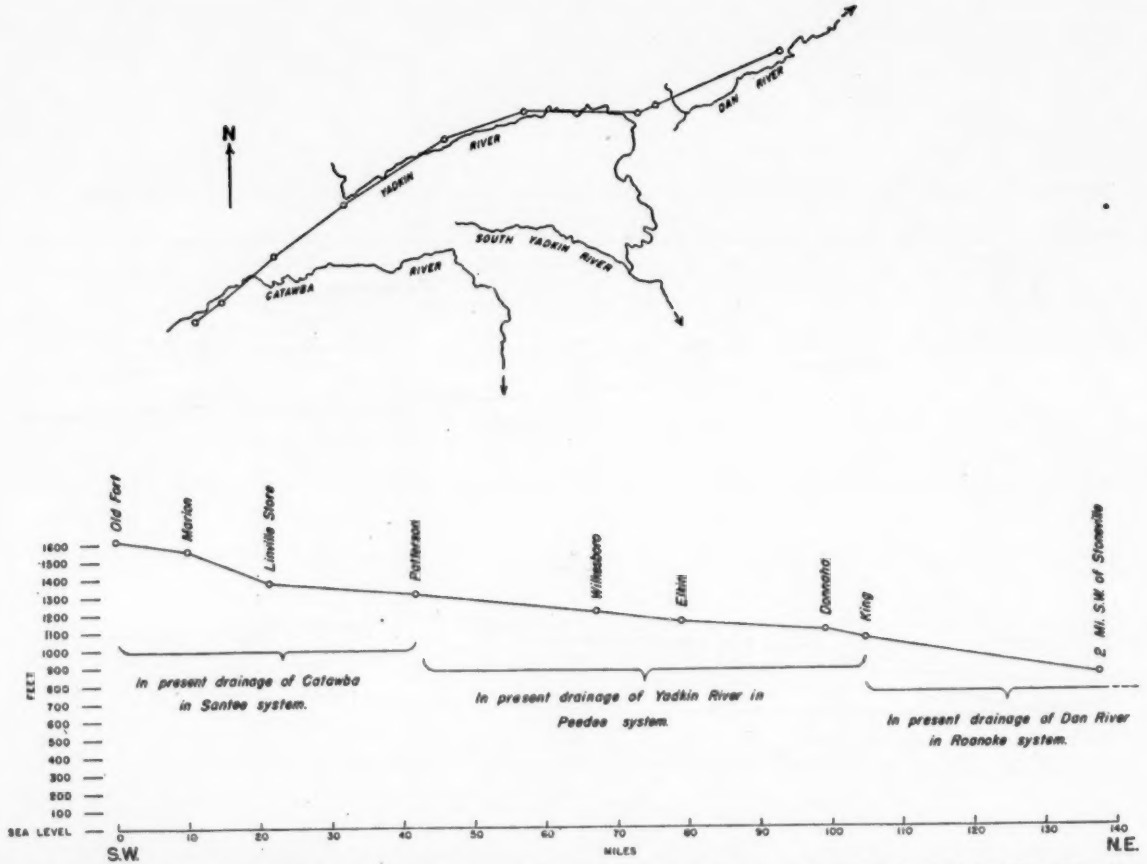


FIGURE 2.—PROFILE ALONG OLD STRATH TERRACE ACROSS PRESENT DRAINAGE AREAS OF CATAWBA, YADKIN AND DAN RIVERS

River T
Cove,
South
Along
kin and
U. S.
noticea
level st
the Pie
and is

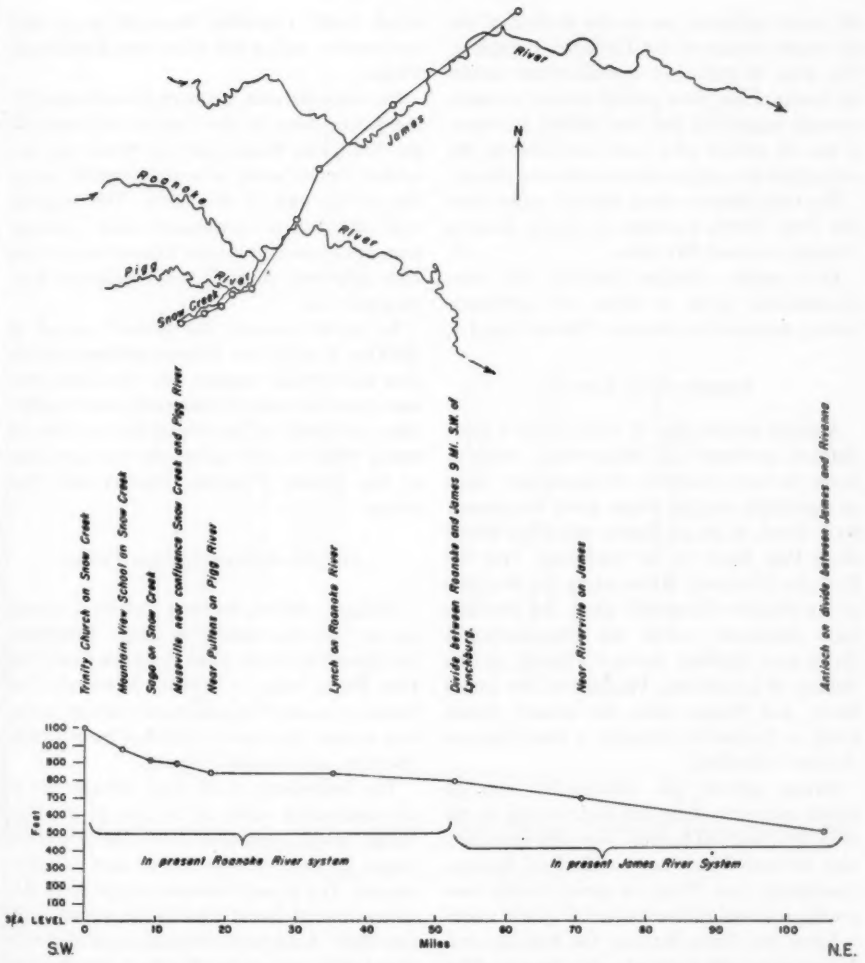


FIGURE 3.—PROFILE ALONG OLD STRATH TERRACE ACROSS PRESENT DRAINAGE AREAS OF SNOW CREEK, PIGG, ROANOKE, AND JAMES RIVERS

Location map above shows profile points.

River Triassic Basin from the vicinity of Walnut Cove, North Carolina, toward Danville and South Boston, Virginia.

Along this axis, the divide between the Yadkin and Dan rivers is essentially the course of U. S. Highway 52. Although there are no noticeable valley walls to locate the old high-level strath where it should cross the divide, the Piedmont Peneplain is well represented and is locally coated with fluvial gravels where the old stream should have crossed it. The gravels can be seen along the highway about a mile northwest of King, North Carolina.

Northeast of Walnut Cove, the Dan occupies a trench in the Piedmont Peneplain, and high-level gravels occur locally on the peneplain surface near by. They can be seen about 2 miles southwest of Stoneville, North Carolina.

A profile connecting the several components of this old fluvial surface is shown in Figure 2.

FIGURE 2.—PROFILE ALONG OLD STRATH TERRACE ACROSS PRESENT DRAINAGE AREAS OF CATAWBA, YADKIN AND DAN RIVERS

All points indicated are on the surface of the old strath terrace or the Piedmont Peneplain. The slope is uniformly northeastward across the basins of the three present stream systems, strongly suggesting that the present remnants of the old surface were once continuous as the valley floor of a single northeast-flowing stream.

The total distance along this old valley from Old Fort, North Carolina, to South Boston, Virginia, is about 200 miles.

Very similar relations between the now-dismembered parts of other old northeast-flowing streams are shown in Figures 3 and 4.

Roanoke-James Terrace

A strath terrace (Fig. 3) which heads a short distance northeast of Martinsville, Virginia, seems to have extended northeastward along an essentially straight course down the present Snow Creek to its confluence with Pigg River; along Pigg River to the confluence with the Roanoke (Staunton) River; along the Roanoke to the vicinity of Leesville where the Roanoke turns southeast; across the Roanoke-James divide near Bedford Springs, Virginia, to the vicinity of Lynchburg, Virginia, on the James River, and thence down the present James River to Scottsville, Virginia, a total distance of about 100 miles.

Stream gravels are occasionally seen on strath remnants along this line, notably in the vicinity of Bedford Springs, along the secondary road between Forest and the Bedford Springs-Lynchburg road. They are spread widely over a well-preserved remnant of a flat upland where it forms the divide between the Roanoke and James rivers. The gravels also appear along other roads in this vicinity and can readily be seen at New London Academy near Bedford Springs.

The gravels occur in the center of a broad strath terrace about 10 miles across, which extends northeastward across the present principal drainage lines. Between the Roanoke and the James, this strath is limited on the northwest by Fleming Mountain and the southern end of No Business Mountain. On the northwest, it is limited by a discontinuous chain of low ridges which rise to elevations of 1000 to 1500 feet (200–700 feet above the strath level). Chandler Mountain is at their northeastern end, a few miles from Lynchburg, Virginia.

For some distance upstream from Scottsville, the James flows in the Triassic sediments of the Scottsville Basin, and the strath just described trends along a southwestward extension of the axis of this basin. This suggests that the former northeast-flowing drainage here was controlled by the Triassic rocks when they extended much farther southwest than they now do.

In similar manner, the present control of the Dan River by the Triassic sediments of the Dan River Basin suggests that those sediments also extended much farther southwest in earlier times and localized the stream that cut the old strath which extends across the drainage areas of the present Catawba, Yadkin and Dan rivers.

Rockfish-Mechum-Rapidan Terrace

Similar relations between high-level straths can be seen in a number of places throughout the upper Piedmont. Thus along the foot of the Blue Ridge Scarp in Nelson, Albemarle, and Greene counties, Virginia, such a surface can be seen to cross the present valleys of the Rockfish, Mechum, and Rapidan rivers.

The headwaters of all these streams are in one continuous valley at the foot of the Blue Ridge Scarp, separated from the Piedmont proper by a now-discontinuous zone of higher ground. The present streams escape from this valley through broad gaps in this eastern wall, but their headwater reaches must formerly have been integrated within the valley, because the dissected remnants of a high-level strath persist across their now-separate drainage areas. This strath descends regularly to the northeast (Fig. 4). There are abundant remnants of gravel-beds visible in the dissected edges of the strath, and there seems little doubt that a single stream formerly flowed northeastward throughout its length.

Appomattox-James Terrace

A similar relation exists between the Appomattox and James rivers (Pl. 1). The Ap-

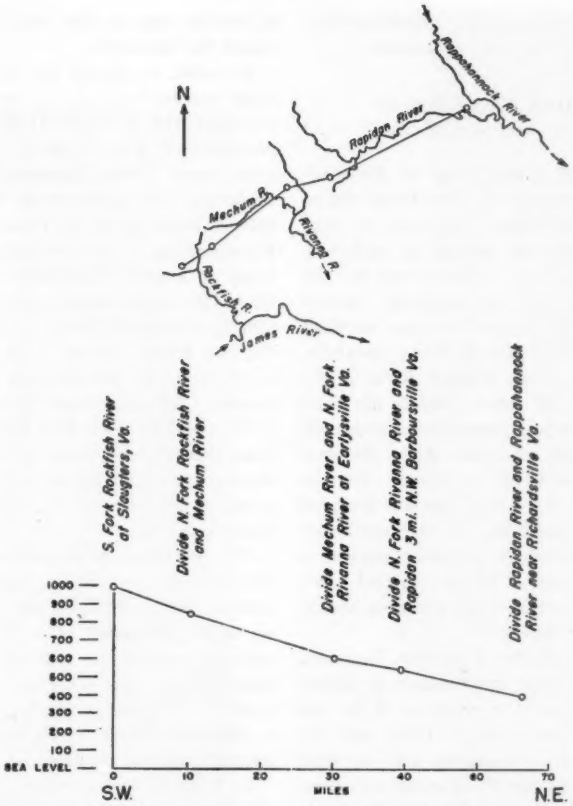


FIGURE 4.—PROFILE ALONG OLD STRATH TERRACE ACROSS PRESENT DRAINAGE AREAS OF ROCKFISH, MECHUM, RIVANNA, AND RAPIDAN RIVERS

Location map above shows profile points.

pomattox flows northeast from the vicinity of Farmville, Virginia; at that point, a tributary, Briery Creek, enters it along the same trend from the southwest. Near Clementown Mills, Virginia, in the northeastern corner of the Farmville sheet, the Appomattox turns off toward the east, and Deep Creek, a small tributary of the James, continues along this same northeast trend beyond the point where it is abandoned by the Appomattox. The headwaters of Deep Creek approach very closely to the main stream of the Appomattox at the point where the latter turns east. Nevertheless, wide, flat remnants of the Piedmont Peneplain are preserved on the divide. Near Sunnyside, on Virginia Highway 13, gravels are preserved on these remnants.

Description of Gravels

The gravel deposits in the region affected by the northeast-flowing ancestral drainage are composed of cobbles ranging in general from 2 to 8 inches in maximum diameter. They vary from subrounded to highly rounded, and do not resemble marine gravels. In many places they are seen in section in road-cuts where they usually are overlain by the silty lateral and vertical accretion deposits typical of floodplains (Pl. 3, fig. 2).

On the flat surfaces of the old straths, such gravels are seen only as a scattering of occasional cobbles on the surface of the ground. But, ordinarily, one would not expect an undissected floodplain surface to show its channel

deposits before the lateral and vertical accretion deposits had been removed by erosion.

Topographic Expression of Former Drainage Ways

Along the northwestern edge of the Piedmont, near the foot of the Blue Ridge Scarp, the old northeast-flowing drainageways tend to follow intermontane valleys. In such constricted locations, it is easier to find the old channel deposits on the relatively narrow strath terraces where they have been trenched by the present drainage. In many instances, as in the case of the Yadkin River above Wilkesboro, the old strath terraces are little wider than the present floodplains, and gravels are fairly continuous along them. In such cases it would seem fair to assume that the greater share of the strath terrace was occupied by the former floodplain of the northeast-flowing drainage, which probably reached a more advanced stage of the erosion cycle than has the present drainage; and therefore should have had a wider floodplain.

Farther down in the Piedmont, however, there are no such constricted valleys to delimit the old straths, and the remnants of the old surface suggest that it was a fairly well developed peneplain, widespread and of little relief. On such a surface there would have been little difference in elevation between the floodplains of the streams and the general level of the upland divide areas. Therefore, after relocation of the drainage and dissection of the surface, it would be somewhat difficult to distinguish the small areas where the floodplains had been from the large interfluvial areas where the general upland had been. For this reason, the continuity of the old northeast-flowing streams across present divides is not as readily seen in the central Piedmont as nearer the mountains.

To assist in tracing the former valleys in these obscure locations, a topographic model was made (Pl. 3, fig. 1). It is composed of 32 cross-sections located along lines spaced 15 miles apart (where topographic maps were available). The northernmost of these sections extends from Carlisle in Cumberland County, Pennsylvania, at the northern end of the Blue Ridge to Aberdeen in Harford County, Maryland. The southernmost extends from Bakersville in Mitchell County, North Carolina, on the Blue Ridge Upland, to the general vicinity of Charlotte, in Mecklenburg County, North Carolina. All sections extend from points west of the crestline of the Blue Ridge Scarp to the inner edge of the Coastal Plain except where data were unavailable for lack of topographic maps, as in much of the lower Piedmont of North Carolina.

The continuity of the strath terraces along the old northeastward drainageways is readily seen on the model itself but it is difficult to see in the photograph. The strath terraces lose elevation consistently with distance northeastward and are commonly bounded by higher land, the remnants of which correlate to reveal a dissected surface which slopes consistently and uniformly to the east.

In Virginia and northern North Carolina, this higher surface is preserved in the remnants of the Piedmont Plateau. It slopes smoothly in an easterly direction, making an angle of some 30° to 40° with the direction of flow of the former northeast-flowing streams.

Since the Piedmont is an area of rather minor relief and the horizontal distances involved in the model amount to as much as 400 miles, it was necessary to use an extreme vertical exaggeration. The actual ratio is approximately 120

Plate 2.—STRATH TERRACES

Figure 1. View from U. S. Highway 70 between Marion and Old Fort, North Carolina
Present floodplain of South Fork of Catawba River in foreground, 1500-foot strath terrace in middle ground, lower spurs of Blue Ridge in background.

Figure 2. Valley of Lower Creek from a point near the intersection of U. S. Highway 321A and North Carolina Highway 18 near Lenoir, North Carolina
Present floodplain in foreground, 1500-foot strath level in middle ground, outliers of Brushy Mountains in background.

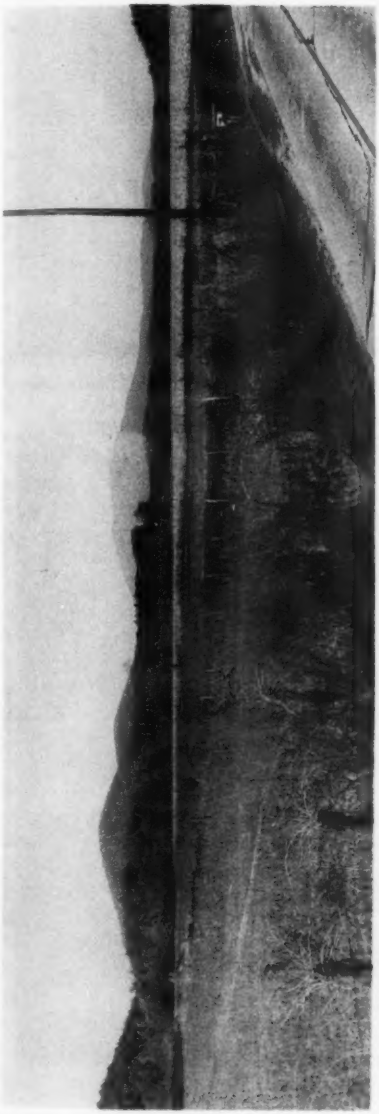


FIGURE 1

FIGURE 2

STRATH TERRACES

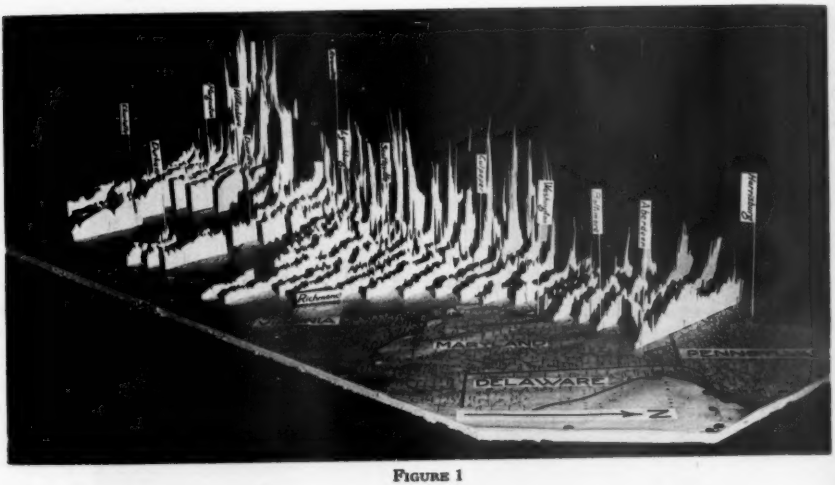


FIGURE 1

FIGURE 2

TOPOGRAPHIC MODEL AND GRAVEL BED

to 1. Yet, the old eastward-sloping surface is clearly seen on the model and shows no great irregularities. Actually it is a surface of marked uniformity much smoother than it appears on the model and well deserves the term peneplain.

The writer thinks this upper surface originally sloped northeast in the direction of the axial streams of the former drainage system, but was tilted up along its northwestern edge at the time of the reactivation of the Blue Ridge border fault (White, 1950).

Such southeastward tilting of a northeast-sloping surface would produce a slope oriented somewhere between the original northeastern slope and the southeastward tilt, or approximately east (Fig. 5).

The seaward slope thus acquired probably accelerated the multiple captures and final dismemberment of the old northeast-flowing drainage which seems to have been indigenous to this older surface.

Actually the orientation of the resultant slope, and that of the transverse trunk streams developed on it, change in conformity with the curve of the Blue Ridge Front, the edge of the Coastal Plain, and the present coastline, all of which veer increasingly to the west with increasing distance southwestward. Thus the axial trends of the Rappahannock, James, and Roanoke in Virginia are more nearly east than the southeasterly trends of the Yadkin-Peedee and the Catawba-Wateree-Santee in the Carolinas.

Further support for the idea that a northeast flowing stream system has been dismembered by east and southeast-flowing streams may be found in the fact that remnants of the older (higher) surface combined to approximate a single eastward-sloping plane as shown by the model, whereas the lower surfaces developed at the present flood-plain levels of the several discrete southeast-flowing stream systems do not correlate or fall into a smooth curve when projected from one stream system to the next. In other words, in topographic profiles across the present drainage, the present flood plains would appear at discordant levels, but the remnants of the old strath terraces would fall largely in one smooth curve.

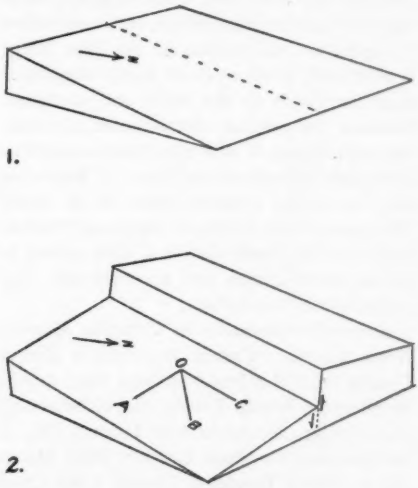


FIGURE 5.—DIAGRAMMATIC SKETCH SHOWING THE GENESIS OF EASTWARD SLOPE OF PIEDMONT SURFACE

(1) Before activation of the Blue Ridge border fault during period of northeast-flowing drainage, regional slope is northeastward. (2) After faulting, downthrown Piedmont block is tilted southeastward parallel with O-A, but regional slope is eastward parallel with O-B because of original northeastward slope parallel with O-C.

Thus it would appear that the upper surface, which can be reconstructed to reveal a single tilted peneplain, was developed in all its parts by a single integrated stream system rather than by the several discrete systems of the present drainage.

PERIPHERAL MONADNOCKS

The extent of the drainage area of the old northeast-flowing streams can be roughly de-

PLATE 3.—TOPOGRAPHIC MODEL AND GRAVEL BED

FIGURE 1. TOPOGRAPHIC MODEL SHOWING PIEDMONT AND BLUE RIDGE SCARP IN NORTH CAROLINA, VIRGINIA, MARYLAND, AND SOUTHERN PENNSYLVANIA

FIGURE 2. GRAVEL BED AT CREST OF VALLEY WALL OF JAMES RIVER, BROWN'S LANDING ABOUT 2 MILES UPSTREAM FROM SCOTTSVILLE, VIRGINIA

lineated by topographic prominences around its edges. Thus the area occupied by the asymmetric tributaries described above can also be delineated by a chain of monadnocks which border it on the south and southeast. The divide at the northwest would originally have been the ancestral Blue Ridge, and, even before its uplift by reactivation of the Blue Ridge Border fault, it was no doubt more conspicuous than the divide at the south and southeast. However the position of this latter, although less conspicuous, is shown by the monadnocks (indicated by hachures on Plate 1). Beginning with the rather extensive mass of the South Mountains in the vicinity of Burke and Rutherford counties, North Carolina, they extend in an arc which curves first eastward and then northeastward as follows:

Andersons Mountain in Catawba County, North Carolina; Youngs Mountain in Rowan County near the Iredell county line; Dunns Mountain in Rowan County east of Salisbury; Flat Swamp Mountain and Jackson Hill in southwestern Davidson County; Pilot Mountain in eastern Randolph County; Cane Creek Mountain in southwestern Alamance County; Hickory Mountain in Chatham County; the Grampian Hills of Orange and Chatham counties, and Nunn, Bald, and Blackwoods mountains of Orange County; Red Mountain in northern Durham County, and Fullers and Hugas mountains near Roxboro in Person County.

There is also some suggestion that this chain of monadnocks may have continued still further northeast at the time the old northeast-flowing stream system was dismembered. The present chain dies out near the North Carolina-Virginia state boundary at the place where it intersects the line which seems to mark the maximum former westward extent of the Coastal Plain (Pl. 1). Assuming that it has been properly located, the monadnocks may well have continued some considerable distance farther northeast before they were submerged and reduced to the general level by wave erosion.

These monadnocks seem to delineate most closely the divide of the old northeast-flowing drainage system; others lie in what seems to have been the same general area of the old divide but actually within the headwater areas of adjacent contemporary drainages. Among these are: Spencer, Pasaur, and Bugger mountains in Gaston County, North Carolina; Kings and Chrowders mountains in Gaston and Cleveland counties, North Carolina, and Cherokee County, South Carolina; Henry Knob in York County, South Carolina, and many monadnocklike prominences of the Uwharie Hills in Randolph and Montgomery counties, North Carolina.

Although many of these hills are inconspicuous, they are the only such hills in the area. In the center of the old drainage area, between this zone of monadnocks and the foothills of the mountains on the northwest, the upper surface of the Piedmont attains its flattest aspect. The writer knows of no hills which rise prominently above it, although several of the more western remnants of the old drainage system flow between the principal front of the Blue Ridge and such outlying masses as the Brushy Mountains of North Carolina.

Profiles along Present Streams

The possibility that this southeastern zone of monadnocks is an old divide area is further supported by profiles of the transverse trunk streams which now drain the region. Three such streams in the Carolinas (Fig. 6), the Catawba-Wateree, the Yadkin-Peedee, and the Rocky, a principal tributary of the Peedee, show a marked steepening of gradient where they cross the Carolina slates, a zone of rocks which seems more resistant to erosion and weathering than are the more coarsely crystalline and generally more feldspathic rocks to the northwest. Perhaps it may be inferred from this steepening that the slates formed a resistant barrier which long protected the old northeast-flowing drainage against capture, allowing it to bring its drainage area well toward peneplaination despite the essentially unstable orientation of its long, arterial streams parallel with the coast. The profiles show that, even after capture, the dismembered portions of the old drainage have much gentler gradients than do the transverse trunks where they cross the slates.

When one considers that a large segment of

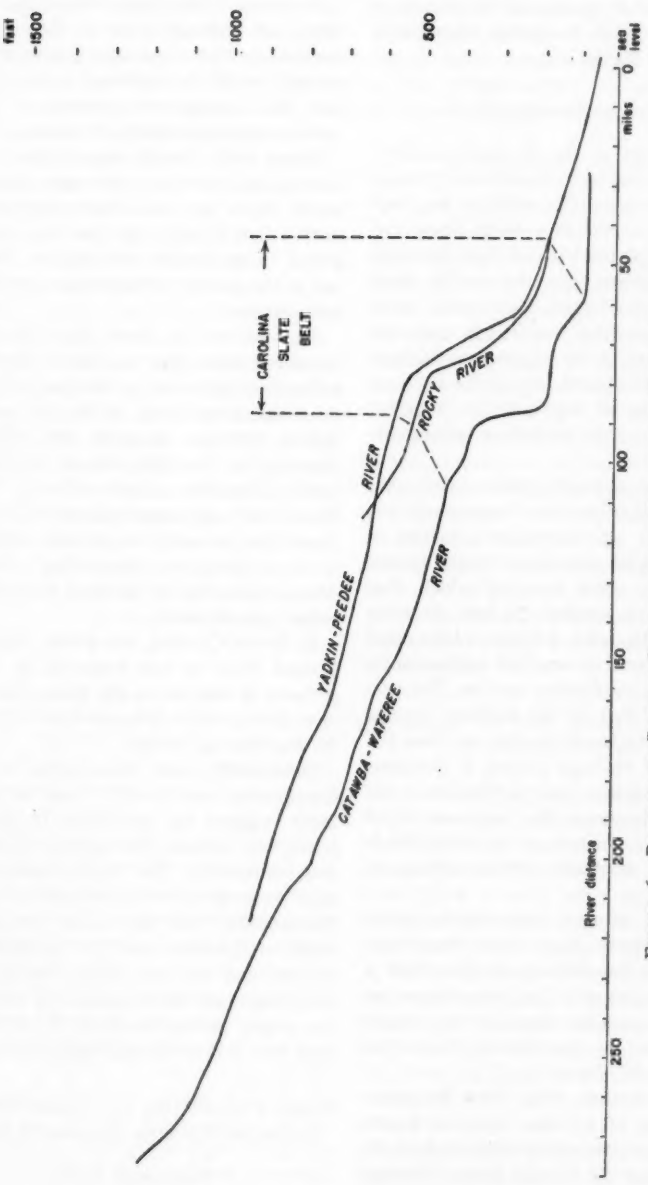


FIGURE 6.—PROFILES OF PRESENT STREAMS ACROSS PIEDMONT OF NORTH CAROLINA
After U. S. Army Engineers in part.

the peripheral monadnock zone lies in the axis of the Slate Belt (Pl. 1) and is coextensive with the zone of steepening in the profiles, it seems plausible that it should represent a former principal divide.

Relation to Coastal Plain

Dismemberment of the old northeast-flowing drainage has not been accomplished through stream capture alone. The width of the Piedmont decreases as the Blue Ridge Scarp and the inner edge of the Coastal Plain converge toward the northeast. Also the zone in which superposition from former coastal plain cover is suggested encroaches increasingly upon the southeastern flank of the old drainage basin as it extends northeastward toward the Potomac River. The zone of superposition is rather clearly marked by the distinctive stream pattern it displays.

The drainage of coastal plains varies considerably with the character of emergence, but a very common type comprises a number of subparallel streams with long, closely spaced, and essentially equal branches which flow parallel with one another for long distances before joining. In such a drainage, a large number of closely spaced streams flow independently to the sea, and confluences are few. The pattern resembles that of the incipient parallel gullies which commonly develop on loose fill.

This type of drainage pattern is dominant in the present coastal plain of Virginia on the west side of Chesapeake Bay, and most of the streams which contribute to it extend headward into the Piedmont without change of pattern.

Furthermore, many of these closely spaced subparallel streams have their headwaters located along a smoothly curved line which is essentially parallel with the present coastline and forms a plausible limit for the former landward extent of the Coastal Plain. (*See* dotted line "S-S", Plate 1.)

The longer streams which have headwater areas northwest of this line would have been those extended consequents which drained the Piedmont across the Coastal Plain. Although these streams share the characteristic subparallel pattern in those parts of their drainage areas which are southeast of the line "S-S", their patterns on the northwest side of the line are quite different.

Northeast of the Neuse River (North Carolina), the drainage areas of these extended consequents tend to broaden greatly and bend around toward the southwest as they incorporate the dismembered reaches of the old northeast-flowing subsequent drainage.

Where such former coastal plain overlap covered any portion of the older drainage, it would have let the dissevered headwater section flow directly into the sea during the period of maximum submergence, becoming one of the present extended consequents upon reemergence.

Southwest of the Neuse River the drainage patterns suggest that maximum submergence and coastal plain overlap fell short of reaching the southeastern flank of the old northeast-flowing drainage. However, the evidence is obscured by the influence of the Durham-Sanford-Wadesboro basin of the Triassic. Here the soft impermeable shales of the Newark Series form an easily eroded belt which seems to have localized a longitudinal subsequent drainage like that of the inner lowlands of the belted coastal plains.

In North Carolina, the former edge of the Coastal Plain as thus suggested by drainage patterns is located in the zone of peripheral monadnocks which delineate the old northeast-flowing drainage system.

Superficially, such interrelation of marine transgression and fluvial changes in the Piedmont suggests the possibility of dating the latter, but actually little use can be made of this relationship. The stream captures could have occurred either before or after the marine transgression, and there is no way of telling which of the many manifest transgressions of the sea was the one which reached inland far enough, and late enough, and regressed in the proper manner to cause the residual effects now seen in the drainage patterns.

Other Possibilities for Establishment of Northeast-Flowing Headwater Reaches

Fault-Angle Valleys

Before the writer concluded that the principle of multiple capture was the most plausible

means by which the present asymmetric drainage could have been produced, he considered several alternative hypotheses. Perhaps the best of these is the possibility that the northeast-flowing headwater reaches of such streams as the Yadkin, Catawba, etc., were originally consequent in fault-angle valleys at the foot of the Blue Ridge Scarp (White, 1950).

Offhand, this would appear the most plausible explanation of a series of independent streams following the same lineal course at the foot of a fault scarp. However, asymmetry of drainage in the Piedmont is not limited to the foot of the Blue Ridge Scarp, but covers the Piedmont generally from there to the former edge of the Coastal Plain.

It would be difficult to explain the drainage pattern of this wide area by late faulting, unless the fault scarps which engendered it had been peneplaned. For although the Blue Ridge Front forms the most prominent topographic scarp in the eastern United States, the Piedmont is an area of singularly little relief.

Also, the persistence of gravel-coated strath terraces across now separate drainage systems necessitates a hypothesis which postulates dismembering a former northeast-flowing drainage.

Differential Headward Erosion

Another hypothesis would explain the asymmetry by differential headward erosion. Assuming that the transverse trunk streams flowed down a regional slope which was not at right angles to the regional strike, those longitudinal subsequent tributaries which flowed in such a direction as to share a component of the regional slope would have an advantage over the opposing tributaries, which would be barbed in their angular relationship to the main stream, and would be flowing in considerable measure opposite to the regional slope.

There are many arguments against such an evolution. In the first place, several of the master transverse trunk streams such as the Catawba and Yadkin rivers have courses perpendicular to the regional strike (Catawba River below Catawba; Yadkin below Donnaha, North Carolina.) Only in Virginia do any of them make appreciable angles with the strike. Moreover, the asymmetry of drainage south of the Potomac is reversed in orientation with respect to the asymmetry north of the Potomac. Again. if the asymmetry were produced by differential headward erosion, the gravel-coated strath terraces should not persist on smooth profiles across the divides between the present separate stream systems.

Superposition From Alluvium

A third alternative is to assume that aggrading streams flowed across the upper Piedmont when faulting was active in the Blue Ridge. Laden with coarse debris from the dissection of an active fault scarp, such streams might have laid down on the then well-preserved Piedmont Peneplain a veneer of coalescent bar plains somewhat similar to the Ogalalla and Arikaree formations of the High Plains.

A scattering of fluvial cobbles on the undissected remnants of the Piedmont Peneplain in widely separate areas is the chief support for this hypothesis. However, most of these are at localities directly in line with northeast-flowing segments of present-day streams such as those described above. A much more general distribution would be necessary to suggest the former existence of a regional outwash from the mountains.

Exhumation Of Old Valleys From Beneath Coastal Plain

A fourth hypothesis is that a northeast-flowing drainage system existed in Cretaceous or Tertiary time and was buried by coastal plain sediments from which the present transverse trunk streams were superposed. After uplift and dissection segments of the former northeast-flowing drainage were exhumed.

This idea seems to have little merit because the maximum former extent of the Coastal Plain since Piedmont drainage was established is evident from the present stream pattern in the lower Piedmont as described above and shown by line "S-S" on Plate 1. Since all former drainage trends were erased by this phase of coastal plain overlap, any older, more extensive overlap should likewise have erased the former trends farther west.

Also the gravels and restored profiles along strath terraces show that the several dissevered

components of the old drainage were formerly coextensive on the Piedmont surface, and therefore their dismemberment must have taken place at that same level. Perforce, the coastal plain cover from which such super position would have been accomplished should have rested on the present Piedmont peneplain. It is improbable that such a coastal plain cover existed, for the Piedmont peneplain, although extensively dissected, still preserves innumerable level remnants which are so large and flat that if any permeable sandy sediment had ever capped them it should still do so. No remnants of such coastal plain cover can be found, and the possibility that it ever existed is small.

Again, if a mature drainage system were buried and later exhumed by major tributaries of streams flowing at right angles to its original direction of flow, it would seem likely that the direction of flow in a large share of the exhumed segments should be reversed.

Superposition From Coast-Parallel Consequents

Another hypothesis using a former coastal plain cover to produce the northeast-flowing reaches of present streams might be based on the fact that a characteristic type of stream in the Atlantic Coastal Plain flows essentially parallel with the coastline.

Examples are the Waccamaw River of North and South Carolina; the Northeast Cape Fear River of North Carolina; and the streams of the eastern shore of Maryland which flow into Chesapeake Bay, such as the Pocomoke, Wicomico, Nanticoke, Choptank, Tuckahoe, Chester, and Elk rivers.

If such streams had been prevalent as major tributaries of coast-perpendicular consequents on a coastal plain cover which extended across the upper Piedmont, their incision following uplift might have produced the asymmetry in question. However, a major reorientation of coastal plain drainage direction would have had to occur since any such systematically northeast-flowing streams were prevalent in the Atlantic Coastal Plain, for practically all streams flowing parallel to the coast in the present Coastal Plain flow in the opposite direction, or southwestward.

Also, such unguided superposition could not account for the nice subsequence of the northeast-flowing reaches of the present Piedmont drainage. As noted above, these are consistently in weak zones and commonly in Triassic troughs.

The accordance of the restored profiles of strath terraces in adjacent stream systems, as well as the presence of fluvial gravels on intersystem divides, also militates against this idea.

Control By Plunging Structures

A final alternative is the possibility that plunging structures guided the development of asymmetric tributaries by making headward erosion easier in a southwesterly direction than in a northeasterly one. Aside from the fact that such structures are not notably present in the area in question, this idea like the others is disproved by the strath terraces and gravel beds which cross the present divides.

Considering the weakness of all these alternative hypotheses, the writer believes multiple capture is the most plausible explanation of the asymmetry of drainage; although it is not proven, it has considerable supporting evidence and few contradictions.

Relation to Drainage North of the Potomac

Although the writer has confined his field work to the Piedmont region of North Carolina and Virginia, he is tempted to suggest that the ancestral northeast-flowing drainage of that region was but the southerly branch of a still larger stream system. Possibly it had its northern headwaters in the southern slopes of the Adirondacks, for it has already been suggested (Fenneman, 1938) that subsequent insinuation of the Mohawk and Black rivers (New York) along the strike of soft paleozoic sediments captured a previous drainage system which flowed from the Adirondacks southward across the present Mohawk Valley into the Susquehanna, which has its present headwaters in the backslope of the Helderberg Scarp.

The stream pattern in the Appalachian plateau of eastern New York and north-central Pennsylvania, and the Ridge and Valley prov

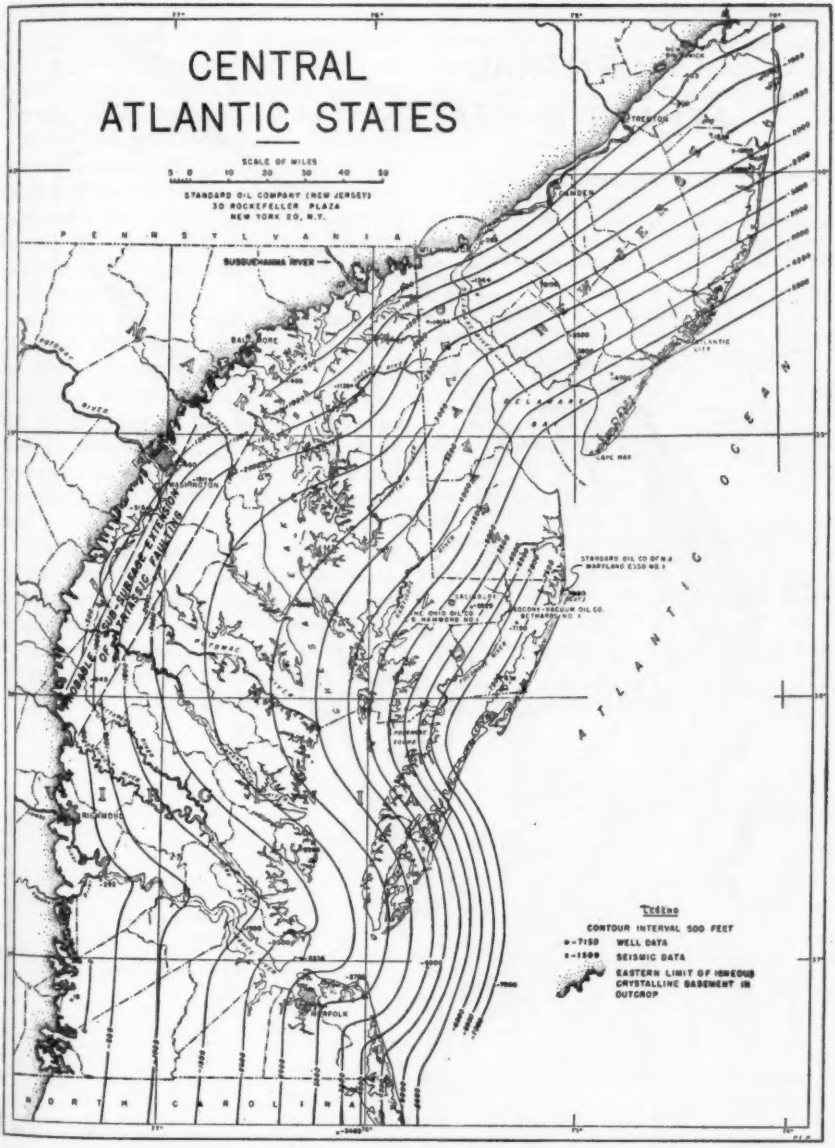


FIGURE 7.—CONFIGURATION OF SURFACE OF BASEMENT ROCKS BENEATH COASTAL PLAIN

Showing embayments opposite points where present Potomac and Susquehanna rivers debouch onto Coastal Plain. (After W. P. Spangler and Jahn J. Peterson, 1950, fig. 12, p. 82.)

ince of Pennsylvania, is in considerable measure the mirror image of the asymmetric pattern of the Piedmont south of the Potomac. Such streams as the Susquehanna and Delaware consistently flow southwestward along the strike and then jog southeastward in transverse

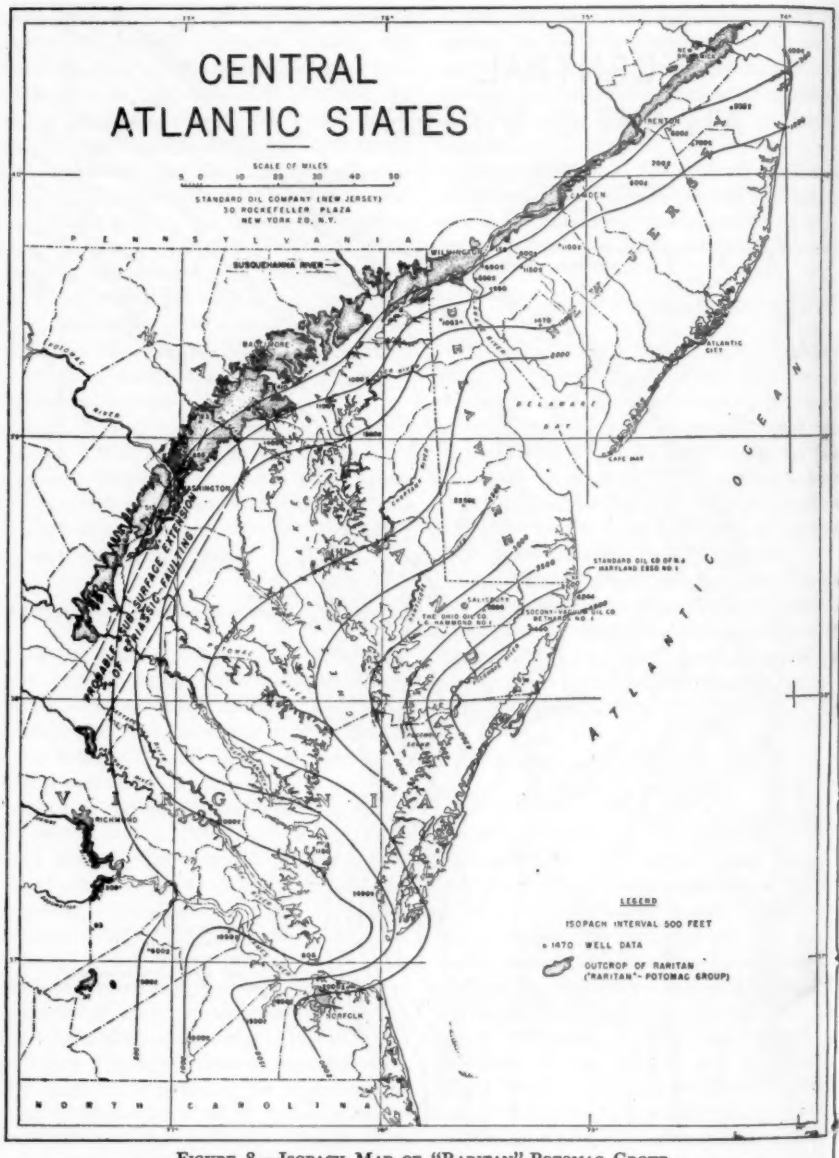


FIGURE 8.—ISOPACH MAP OF "RARITAN"-POTOMAC GROUP

Showing thickening opposite points where present Potomac and Susquehanna rivers debouch onto Coastal Plain. (After W. P. Spangler and Jahn J. Peterson, 1950, fig. 19, p. 89.)

reaches only to turn their courses again southwestward parallel with the strike. However, asymmetry of tributaries is not as noticeable here as in the Piedmont south of the Potomac.

Perhaps some support of the former integration of drainage from north and south may be had from the work of Spangler and Peterson (1950) whose maps show the configuration o

the sub-Cretaceous basement and structure contours and isopachs for the principal formations of the Cretaceous and Tertiary. Their map showing the configuration of the basement (Fig. 7), and an isopach map of the Raritan formation (the basal formation of the Coastal Plain) (Fig. 8) both show embayments opposite the place where the Potomac and Susquehanna rivers enter the Coastal Plain. These are the only two major streams of the present drainage which should have been arterial components of the ancestral drainage.

No similar embayments appear opposite the points where other important streams of the present drainage, such as the Delaware, James, and Roanoke enter the Coastal Plain. The isopach map of the Raritan formation shows a thickening over the embayments in the basement, obviating the possibility that they were formed by post-Cretaceous deformation.

These facts suggest not only that the Potomac and Susquehanna rivers were extant essentially in their present locations in Cretaceous time, but also that they then received a larger share of the regional drainage than they do now.

Further support for the idea of a broad regional integration of drainage can be found in the coincidental fact that the Susquehanna system, the Potomac system, and the restored northeast-flowing Piedmont system all converge toward the head of Chesapeake Bay, the largest estuary of the Atlantic coast.

References Cited

Carlston, C. W. (1946) *Appalachian drainage and the highland border sediments of the Newark series*, Geol. Soc. Am., Bull., vol. 57, p. 997–1032.

Fenneman, N. M. (1938) *Physiography of eastern United States*, McGraw-Hill, p. 39.

Johnson, D. W. (1931) *Stream sculpture along the Atlantic slope*, Columbia Univ. Press.

Spangler, W. B., and Peterson, J. J. (1950) *Geology of Atlantic Coastal Plain in New Jersey, Delaware, Maryland, and Virginia*, Am. Assoc. Petrol. Geol., Bull., vol. 34, p. 1–99.

United States Army Engineers (1933a) *Yadkin-Peedee River, North Carolina and South Carolina*, House of Representatives Doc. no. 68, U. S. Gov. Printing Office, pls. 2, 3, 4.

—— (1933b) *Santee River, North Carolina and South Carolina*, House of Representatives Doc. no. 96, U. S. Gov. Printing office, pl. 6.

White, W. A. (1950) *The Blue Ridge Front, a fault scarp*, Geol. Soc. Am., Bull., vol. 61, no. 12, pt. 1, p. 1309–1346.

University of North Carolina, Chapel Hill, N. C.

Manuscript Received by the Secretary of the Society, January 12, 1951

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 64, PP. 581–598, 6 FIGS., 1 PL. MAY 1953

PLEISTOCENE GEOLOGY OF PART OF CENTRAL SOUTH DAKOTA[1]

BY DWIGHT R. CRANDELL

ABSTRACT

The area around Pierre, South Dakota, is bisected by the trench of the Missouri River, which forms a boundary between a region mostly covered by glacial deposits that overlie Pierre shale on the east and a region directly underlain by shale on the west.

The earliest Pleistocene deposits are remnants of stream alluvium derived from the region to the west. These are now perched on some of the highest knobs in the area and bear witness to an inversion of topography caused, in part, by a lowering of base level when the Missouri River trench was formed marginal to the glacier of the Illinoian age.

In the Wisconsin age, the glacier of the Iowan sub-age crossed the Missouri trench and in one place extended 20 miles farther west. The westward extension of each successive glacier was less than its predecessor. The glacier of the Tazewell sub-age reached the Missouri River, and in the course of its retreat built two massive end moraines in the area. Deposits of this substage cover most of the area east of the Missouri. Till of the Cary substage is restricted to the northeast corner of the area; the drift border of the glacier of the Mankato sub-age lies about 15 miles east of the area investigated in detail.

Glacial drift of the several substages of the Wisconsin is locally mantled with as much as 30 feet of loess. The Tazewell-Cary interval is represented in this loess by a buried, immature soil profile.

CONTENTS

TEXT

	Page
Abstract	581
Introduction	581
Acknowledgments	583
Pleistocene geology	583
Pre-Wisconsin stratigraphy and inferred history	583
General statement	583
Pre-Yarmouth nonglacial deposit	583
Post-Kansan deposits	583
Pre-Illinoian topography and drainage	585
Origin of the Missouri River trench	587
Sangamon erosion	588
Wisconsin stratigraphy and inferred history	589
Iowan(?) and Tazewell(?) drift	589
Till and moraines	589
Stratified drift	590
Drift borders	591
Iowan-Tazewell interval	591
Tazewell-Cary interval	591
Cary(?) glacial drift	593
Till and moraines	593
Stratified drift	593
Drift border	594
Cary-Mankato interval	594
Mankato glacial drift	594
Wisconsin loess deposits	595
Summary of Pleistocene history	596
References cited	597

ILLUSTRATIONS

Figure	Page
1. Index map of central South Dakota	582
2. Location map of the Pierre area	584
3. Diagrammatic reconstruction of the southern half of the Oahe quadrangle	586
4. Reconstruction of inferred topography and drainage immediately before formation of the Missouri River trench	587
5. Wisconsin ice margins in central South Dakota	592
6. Idealized cross section of the Missouri River trench in the Pierre area	593

Plate	Facing page
1. Generalized geologic map of the Oahe, Pierre, and Canning quadrangles	581

INTRODUCTION

The part of central South Dakota discussed in this paper is an area of about 750 square miles mapped in detail, and an additional surrounding region of about 1500 square miles mapped or visited in reconnaissance, in the field seasons of 1949 and 1950 (Fig. 1). This region is divided by the Missouri River into

[1] This paper is a condensation of part of a dissertation presented for the degree of Doctor of Philosophy in Yale University.

nearly equal parts that are geologically dissimilar. The part west of the river is underlain predominantly by members of the Upper Cretaceous Pierre shale; the part east of the river is underlain by thick deposits of several glacial substages that overlie Pierre shale.

The area mapped in detail includes three 15-minute quadrangles, from west to east: Oahe, Pierre, and Canning (Fig. 1). Geologic mapping included a subdivision of bedrock units as well as of surficial deposits (Crandell, 1950, 1951), but the bedrock stratigraphy is not discussed here.

One purpose of the investigation was to gain a better knowledge of the Pleistocene history of the Missouri River trench, and to relate it to the history of the region west of the Missouri. This paper presents only the most essential field evidence and the general framework of reasoning therefrom, and summarizes some of the more important conclusions reached from the study of the Pleistocene deposits. A more fully documented discussion of the Pleistocene geology will be presented later in a publication of the U. S. Geological Survey.

Correlation of glacial drift in the Pierre area with subdivisions of the standard Pleistocene section have been made possible by Flint's

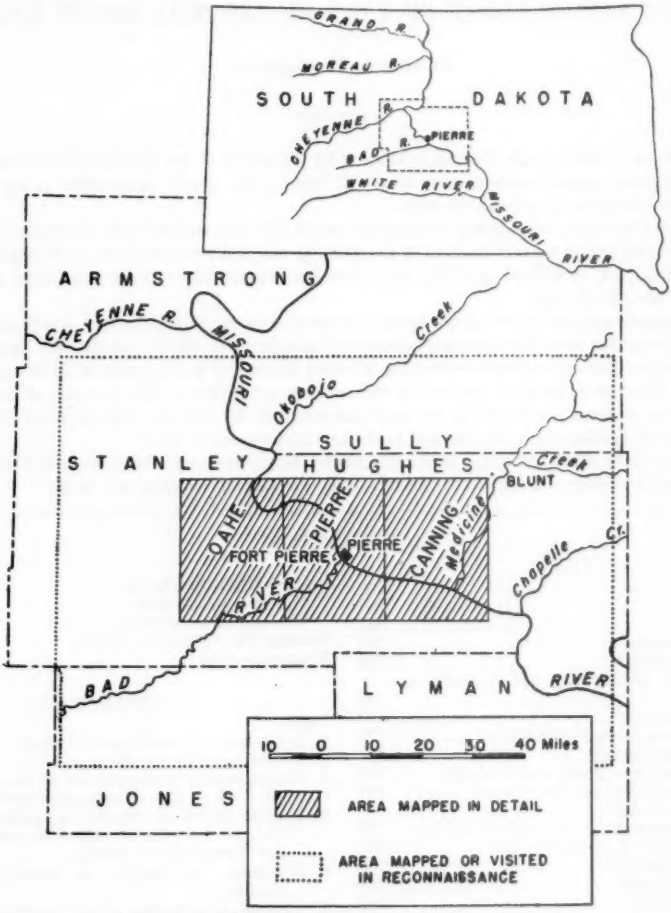


FIGURE 1.—INDEX MAP OF CENTRAL SOUTH DAKOTA

(manuscript in preparation) investigation of the Pleistocene geology of eastern South Dakota.

Acknowledgments

The writer expresses his appreciation to Professor Richard Foster Flint, Yale University, for many helpful suggestions and constructive criticisms made throughout the course of the field work, the laboratory investigation, and the writing of the dissertation. In addition to supervising the problem, Mr. Flint has critically read both the full dissertation and this condensation. Professor Claude W. Hibbard, of the University of Michigan, and Dr. C. Bertrand Schultz, of the University of Nebraska State Museum, kindly identified the Pleistocene vertebrate fossils.

Pleistocene Geology

The pre-Wisconsin Pleistocene history of the region west of the Missouri River is largely one of stream erosion and deposition, controlled at certain times by events in the glaciated region to the east. Evidence pertaining to events before the Wisconsin east of the Missouri River is very scanty, owing partly to erosion and partly to the thick blanket of Wisconsin drift. In contrast, the Wisconsin history is well recorded by the thick glacial deposits in the Missouri River trench and the area east of it, but is imperfectly known in the region west of the river.

The standard Pleistocene sequence used in this report is as follows:

Epoch	Age[1]	Sub-age
Pleistocene	Wisconsin	Mankato Cary Tazewell Iowan
	Sangamon	
	Illinoian	
	Yarmouth	
	Kansan	
	Aftonian	
	Nebraskan	

[1] The official classification and nomenclature of the U. S. Geological Survey differs in some respects from those used in the present paper.

Pre-Wisconsin Stratigraphy and Inferred History

General Statement

No pre-Wisconsin glacial drift has been identified in the Pierre area. If ever present, it is very doubtful that it survived the extensive dissection that followed initial cutting of the Missouri River trench. The oldest known Pleistocene sediments near Pierre are deposits of sand and gravel that cap high knobs west of the Missouri River and are remnants of once widespread bodies of alluvium that formerly extended far east of the present outcrop.

Pre-Yarmouth Nonglacial Deposit

The highest and therefore probably the oldest post-Cretaceous sediment in the Pierre area is a deposit of sand and pebble gravel that caps Standing Buttes (Fig. 2). The top of the highest butte lies at an altitude of about 2215 feet, and is 8 miles west of and about 780 feet above the valley floor of the Missouri River. The pebbles consist of carbonate rocks and chalcedony that can be matched with formations of Cenozoic age in west-central South Dakota, indicating western provenance.

The Standing Buttes deposit contains bone fragments and teeth of mammals. Dr. Hibbard tentatively identified one specimen as a fragment of a lower molar of *Hippotigris* (*Plesippus*). Bones of this horse are found in deposits ranging from upper Pliocene to lower Yarmouth. Because the tooth is abraded, it is possible that it was derived from a deposit that antedates the sediments at Standing Buttes.

The presence of bones of terrestrial mammals indicates that the deposit at Standing Buttes was formed by a stream, and the lithology of the pebbles indicates that this stream must have come from the west. The stream probably left a band of alluvium along its course east or northeast of Standing Buttes, but this alluvium has been completely removed by erosion, or has been buried under Wisconsin drift.

Post-Kansan Deposits

Capping Willow Creek Butte (Fig. 2) is a deposit of sand and pebble gravel at an altitude

of about 2090 feet, or about 650 feet above the Missouri River floodplain 5 miles north of the butte. The deposit is similar lithologically to the deposit at Standing Buttes.

these specimens were identified by Dr. Hibbard as "*Equus* and compare well with what I call *Equus niobrarensis*. The range of this type of tooth has been found in the High Plains from

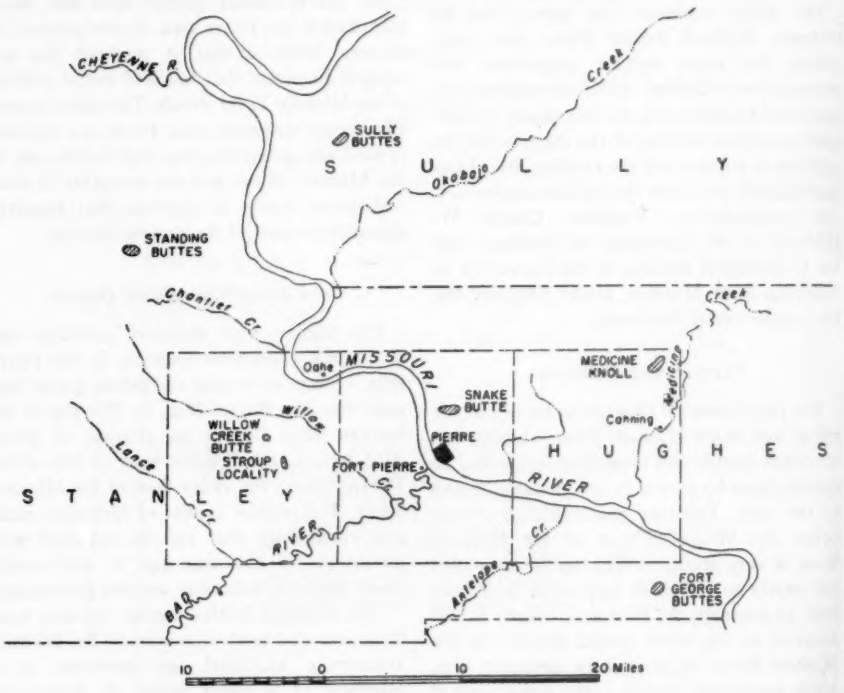


FIGURE 2.—LOCATION MAP OF THE PIERRE AREA

South and east of Willow Creek Butte are extensive deposits of sand and gravel at an altitude of about 2000 feet, about 90 feet lower than the top of Willow Creek Butte. The largest of these will be referred to informally as the Stroup deposit (Fig. 2). This deposit also lithologically resembles that of Standing Buttes. Sand sizes predominate, but small cobbles are common and a few large cobbles are present. The sediment is poorly stratified, with local poorly developed cut-and-fill stratification.

A large number of bone fragments and teeth have been found in the Stroup deposit. Many are so little abraded that they do not appear to have been transported far by a stream. Accordingly, it seems likely that the fossils were not derived from an older deposit. Some of

just below the base of the Pearlette ash to the top of the Kingsdown." This statement indicates that the horse teeth collected have been found elsewhere in deposits ranging in age from early Yarmouth to Wisconsin. Dr. Schultz identified these and other specimens from the same locality as teeth and bones derived from *Equus excelsus*, ref., and stated that they "appear to have come from post-Kansan deposits (Sappa [late Kansan or early Yarmouth] or later), possibly as late as Illinoian."

Thus Hibbard and Schultz agree that the bones and teeth from the Stroup locality are from post-Kansan horses. However, since the stratigraphic dating of the horses is based on occurrences in sediments that overlie Kansan

till, which is itself unfossiliferous, this fauna could be as old as late Kansan.

The Stroup deposit is but one of a number of similar bodies of sand and gravel north of the Bad River at altitudes of 1800 to 2000 feet. These deposits seem to be remnants of a northeast-trending band of coarse floodplain alluvium, deposited when the Bad River was flowing on a surface 300 to 500 feet higher than the present valley floor. No comparable deposits of alluvium have been found south of the Bad River, which suggests that the ancestral Bad River progressively shifted southward after it deposited the alluvium that now caps Willow Creek Butte (Fig. 3).

Pre-Illinoian Topography and Drainage

The presence of stream alluvium capping some of the highest hills west of the Missouri River in the Pierre area demonstrates that there has been topographic inversion since the alluvium was deposited. Although there is no evidence of previous topographic inversion between the times of formation of the Stroup and Standing Buttes deposits, this interval might have been as long as the interval between the date of the Stroup deposit and the present. The relatively greater amount of erosion since this deposit was formed was caused by cutting of the Missouri River trench and the resulting lower base level. Broad gentle uplift of the Great Plains in the last two ages of the Pleistocene epoch may have contributed to the quickening of erosion.

Snake Butte and Medicine Knoll (Fig. 2), two prominent ridges east of the Missouri River in the Pierre area, are collinear with the northeast-trending band of high-level alluvium west of the Missouri and are inferred to coincide with the course of the ancestral Bad River. Although they are now drift-covered, these ridges probably owe their height to a gravel cap similar to that on the high areas west of the Missouri River. Likewise it is inferred that the Sully Buttes in western Sully County represent a ridge once protected by a cap of sand and gravel that might have been an eastward continuation of the alluvium on Standing Buttes, and which might have been the course of the ancestral Cheyenne River.

It seems likely that, in pre-Yarmouth time, two major east- or northeast-flowing streams, the ancestors of the present Bad and Cheyenne rivers, crossed what now is the Missouri River trench and continued eastward. The location of their upstream and downstream courses is not known, but it is possible that both streams headed as far west as the Black Hills.

In the Yarmouth(?), the ancestral Bad River cut down and planed laterally southward, leaving protective caps of sand and gravel on wide areas of its former floodplains (Fig. 3). These caps were more resistant to erosion than the adjacent interfluves underlain only by Pierre shale, and therefore the topography was gradually inverted. Standing Buttes, originally part of a valley floor, now is the highest point on the divide between the Bad and Cheyenne drainage basins in the Pierre area. The inferred band of stream alluvium that trended eastward from Willow Creek Butte to Snake Butte and Medicine Knoll became a high divide between the ancestral Bad River and one of its tributaries. The present southern edge of the Bad River basin may have appeared at this time. This divide now is marked by high ground, 18 miles south of Pierre, that trends northeast and joins the Fort George Buttes (Fig. 4).

The name *ancient Bad River* is here used to refer to a former stream, which, immediately before the cutting of the Missouri trench, occupied nearly the same geographic position but flowed on a higher profile than the present Bad River. It flowed, however, eastward beyond the present mouth of the Bad River, through Hughes County, into what now is the James Valley lowland (Flint, 1949). In crossing the course of the Missouri and continuing eastward, the ancient Bad River was analogous to its predecessor that deposited the alluvium at the Stroup locality and at Willow Creek Butte.

The exact course of the ancient Bad River beyond the mouth of the present Bad is not known with certainty, but it is believed to coincide with a broad linear sag on the upland east of Pierre that is collinear with the trench of the Bad River (Fig. 4). The altitude of the valley floor of the ancient Bad River is believed to have been about 1750 feet, or about 300 feet above the present Missouri River.

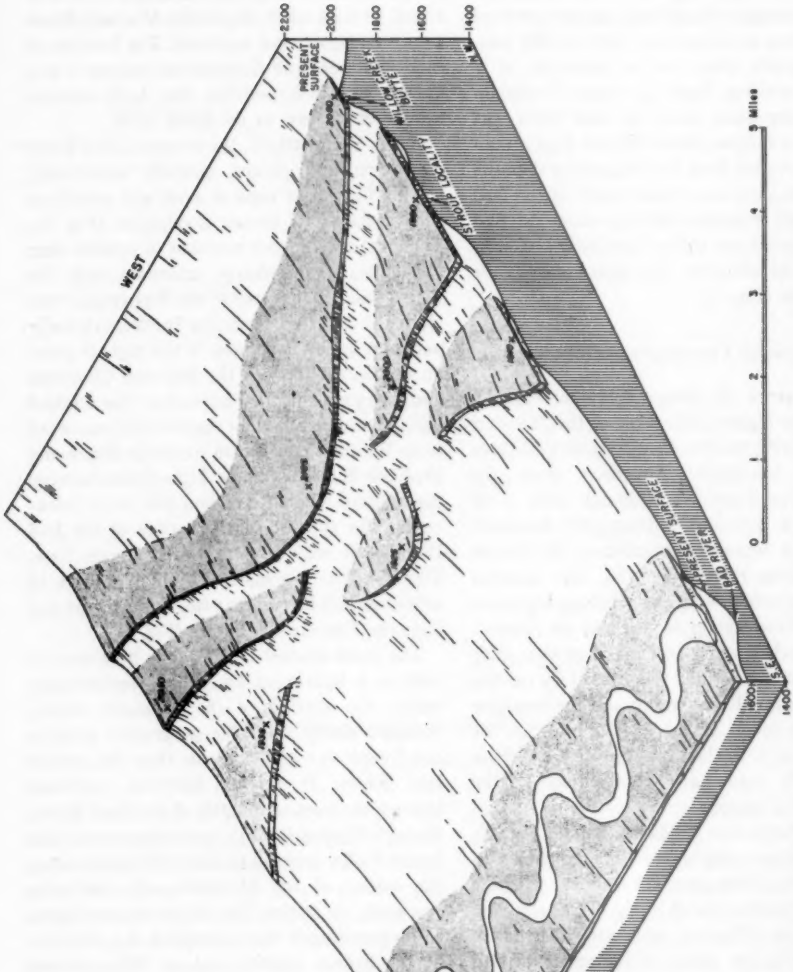


FIGURE 3.—DIAGRAMMATIC RECONSTRUCTION OF THE SOUTHERN HALF OF THE OAHE QUADRANGLE

Showing the probable origin of bodies of alluvium derived from the west and deposited as bands of sand and gravel on Pierre shale straths during lateral planation and downcutting by the ancestral Bad River. Probably the original terrace scarps were never as sharp and steep as shown here. The Bad River is flowing on a fill graded to the Iowan(?) ice sheet or to Iowan(?) outwash in the Missouri River trench. Altitudes and locations on the figure indicate existing

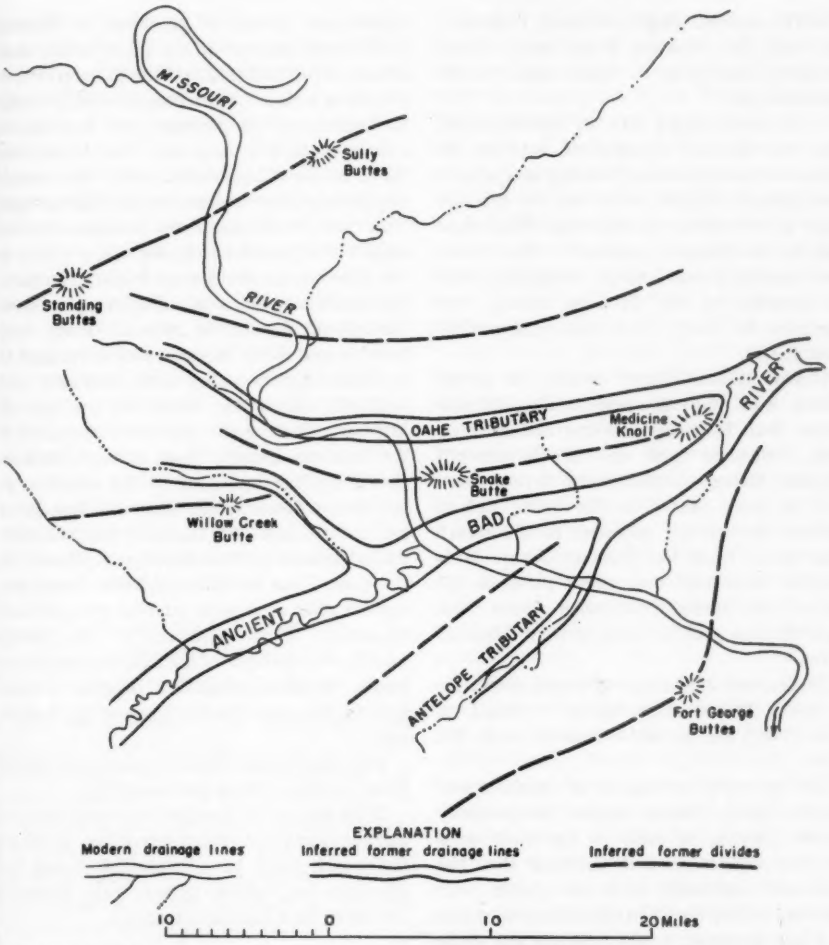


FIGURE 4.—RECONSTRUCTION OF INFERRED TOPOGRAPHY IMMEDIATELY BEFORE FORMATION OF THE MISSOURI RIVER TRENCH

Reconstruction in Sully County in part after Flint (manuscript in preparation).

Origin of the Missouri River Trench

The Missouri River trench is by far the most impressive topographic feature in the Pierre area. The most recent discussions of the time and mode of origin of this feature are by Flint (1949) and Warren (1952) and many of the facts and inferences discussed below have been previously presented by these writers.

Several anomalous relations of the Missouri River trench to the landforms of the region through which it cuts indicate a major drainage readjustment when the trench was first formed.

1. The pattern of drainage west of the Missouri River is conspicuously different from the pattern east of the river. Five large east-flowing streams enter the Missouri in South Dakota (Fig. 1); each of these has a large number of tributaries. East of the Missouri the drainage is poorly integrated; tributaries are less well developed and fewer. Thus, the general drainage pattern in the State is lopsided.

2. The regional slope of South Dakota is eastward. The Missouri River trench trends southerly, nearly at a right angle to the regional slope.

3. At many places east of the Missouri, there are distinct topographic sags in the upland surface, generally trending east, and in some places collinear with one of the five major rivers west of the Missouri. That these sags are downstream segments of the former east-trending stream valleys, abandoned after the creation of the Missouri trench, was suggested by Todd (1914) and confirmed by Flint (1949).

4. The Missouri River trench, in several places, cuts through east-trending divides rather than following a course around their ends, indicating that the trench did not originate through normal stream development.

5. In many places in the State, upland surfaces immediately adjacent to the trench slope away from the Missouri River. It is inferred from this that these parts of the upland were formed by a stream system other than the one graded to the present Missouri River.

The general mechanism of trench formation in South Dakota that follows is based on Flint (1949) and on oral discussions with Mr. Flint.

The westward advance of an ice sheet into central South Dakota caused progressively deeper ponding in each of the main east-trending valleys, as both drainage from the west and meltwater from the glacier were dammed behind the ice barrier. When the water of a lake so formed rose to the altitude of the lowest divide between that lake and an adjacent drainage basin, the lake drained through that outlet and established a channel in the easily eroded Pierre shale. In some localities, the lowest escape route for drainage was along the margin of the ice. By the time the ice margin had retreated to the northeast, a system of connecting channel segments had formed competent to carry both meltwater and drainage from the west. The former eastward extensions of the major river valleys were filled with glacial drift and left in a hanging relationship to the newly formed Missouri River trench.

The best evidence for dating the drainage rearrangement during which the Missouri trench was formed is provided by Warren (1952) from relations in the Chamberlain area (Fig. 1). Warren found that the Missouri River postdates a deposit containing vertebrate fossils of Kansan age or younger, and in addition antedates the Wisconsin age. Thus he inferred that, in the Chamberlain area, the trench originated at the maximum of the Illinoian age.

In the Pierre area, the youngest known deposit that antedates cutting of the trench is the alluvium at the Stroup locality. Surfaces topographically below this alluvium, present on the upland west of the Missouri River, and possibly east of the Missouri also, are graded to a former stream valley that intersects and continues across the trench of the present Missouri. There are no such surfaces graded to the Missouri trench. These surfaces indicate that following deposition of the alluvium at the Stroup locality, the ancestral Bad River cut down at least 100 feet, and that smoothly graded upland surfaces developed adjacent to the river. Thus the Missouri River trench was created after an erosion interval that followed deposition of the alluvium at the Stroup locality. On the basis of the included vertebrate fossils, the Stroup deposit is believed to have been formed after the maximum of the Kansan age.

The oldest known deposit within the Missouri River trench at Pierre is Iowan(?) till.

If the ages of the youngest pre-trench deposit and the oldest post-trench deposit are accepted, the trench must have originated during the Illinoian age, which agrees with Warren's dating in the Chamberlain area.

Sangamon Erosion

Deposits of the Sangamon stage have not been recognized in the Pierre area. This is expectable because Sangamon time was characterized by stream erosion and mass-wasting, quickened by the lower base level created by the cutting of the Missouri River trench.

Stratified drift of the Tazewell(?) substage was deposited in the trench after it had attained nearly its present width and perhaps also nearly its present depth. The lack of well-developed weathering profiles on Iowan drift sheets elsewhere in South Dakota (Flint, in course of publication) indicates that the Iowan and Tazewell interval was a relatively in-

conspicuous event, probably not long enough to accomplish the extensive pre-Tazewell(?) widening of the Missouri trench. Therefore most of the deepening and widening of the trench must have taken place in pre-Wisconsin time.

Logs of wells in the area north of Snake Butte indicate a broad depression in the bedrock surface, from 200 to 300 feet lower than the inferred altitude of the ancient Bad River. There is only one topographic feature low enough to accommodate drainage from this broad lowland: the Missouri River trench. Thus it is evident that the lowland was eroded by a stream or streams graded to the Missouri River. The lowland is now filled with Wisconsin drift of two substages, the Iowan(?) and the Tazewell(?). Because the erosion occurred after the initial cutting of the Missouri River trench and before the Wisconsin, it is referred to the Sangamon age, which must have been of long duration in order to accomplish the amount of dissection indicated.

Wisconsin Stratigraphy and Inferred History

Iowan(?) and Tazewell(?) Drift

Till and moraines.—Two superposed tills, separated by stratified drift, are exposed at two localities near Pierre (test pit exposure and airport exposure, Pl. 1). No pre-Wisconsin till is believed to occur in these exposures, because the base of the section at each locality lies below the altitude of what is inferred to be the pre-Sangamon surface of the area. The two sections are briefly described below:

Test pit exposure

		Feet
7. Loess		3
6. Loess, humified		1.3
5. Loess		5
4. Till		35
3. Stratified drift		12
2. Till		80
1. Pierre shale	More than	2

Airport exposure

		Feet
6. Loess		8
5. Stratified drift	More than	2
4. Till		23
3. Stratified drift		26
2. Till		15
1. Pierre shale	More than	10

Because of the lack of any positive evidence in either exposure to indicate a long period of weathering or erosion, the deposits of each are referred to substages of the Wisconsin stage. Both exposures lie more than 15 miles west of the border of the Cary drift, so the two tills are tentatively referred to the Iowan and Tazewell substages. The presence of two tills in superposition, however, is not necessarily proof that the tills are not of the same age. Either the Iowan glacier or the Tazewell glacier alone could be responsible for the pre-Cary glacial features of the area.

East of the Missouri River, the upland surface is underlain by glacial drift consisting predominantly of clay-rich till. Within a mile or two of the trench, the drift surface is nearly featureless because of the leveling effect of a loess cover that ranges from less than a foot to 20 feet thick. Farther from the river, the drift has a local relief of about 10 feet despite a loess cover as much as 10 feet thick. The surface of the till consists of gentle swales and swells and local closed depressions. This area of low relief underlain by glacial drift is interpreted as ground moraine.

A broad low ridge, 1–2 miles wide, 20 to 60 feet above the surrounding ground moraine, lies about 5 miles northeast of Pierre. This relief feature is underlain by glacial drift and is characterized by a larger number of closed depressions and steeper slopes than occur in the adjacent ground moraine. Because of its ridge-like form and its relatively greater amount of constructional topography than the surrounding ground moraine, this broad ridge is interpreted as an end moraine, and is informally referred to here as the Pierre end moraine (Pl. 1).

Although positive correlation has not been established, the drift of the Pierre end moraine is believed to be essentially the same age as the upper till in the test pit and the airport exposures. The till directly beneath the surficial loess has been correlated by lithology and stratigraphic position from one road cut to another from the airport exposure into the end moraine. On this basis, the end moraine is here referred to the Tazewell(?) substage.

From 1 to 3 miles northeast of the Pierre end moraine and separated from it by ground moraine is a second end moraine, informally referred to here as the Body Ranch end moraine (Pl. 1), 1–3 miles wide and 20–100 feet higher

than the flanking ground moraine. This broad moraine has the general form of a large arc concave toward the west.

The age of the Body Ranch end moraine is less certain than the age of the Pierre end moraine. The till in both moraines is megascopically identical and the two moraines are separated by only 1–3 miles of ground moraine. Also the general parallelism of the two moraines suggests that they were formed at successive margins of the same glacier. The Body Ranch end moraine is more massive and has a larger number of closed depressions, but these characteristics alone probably do not warrant age differentiation of the two moraines. Accordingly, inasmuch as it is several miles west of the Cary drift border, the Body Ranch end moraine also is referred to the Tazewell(?) substage.

Stratified drift.—Although there is no positive evidence that any individual outwash deposit within the Missouri trench is of Iowan rather than Tazewell age, stratified drift deposits of the two substages have been tentatively separated on the basis of relative degree of dissection, grain-size distribution, and altitude. Outwash deposits of the two substages have not been seen in direct contact.

The glacial sand-and-gravel deposits in the Missouri River trench assigned to the Iowan(?) substage have several features in common: (a) The average grain size is generally coarse, and large numbers of cobbles and boulders are included. (b) The deposits are restricted to altitudes ranging from 1800 feet 5 miles north of Pierre to about 1690 feet downstream in the southern part of the Canning quadrangle. (c) Dissection of the deposits has been extensive. From altitudes and character of the deposits it is inferred that Iowan(?) ice-contact stratified drift was deposited along the sides of the Missouri trench in the vicinity of Peoria Bottom (Pl. 1) and that, upon melting of the ice, a valley fill of outwash sand and gravel was built up to an altitude of about 1700 feet at Pierre. Since the pre-Wisconsin bedrock floor of the Missouri trench was at least as low as 1500 feet, the Iowan(?) outwash fill must have been at least 200 feet thick.

Remnants of sand and gravel in the Bad River trench (Pl. 1) that range in altitude from 1800 feet at the west margin of the Oahe quadrangle to 1710 feet near Fort Pierre represent a former valley fill that evidently was graded to a fill in the Missouri River trench. The only known fill in the Missouri that reached this altitude is Iowan(?). Accordingly the 1710- to 1800-foot fill in the Bad River valley is inferred to be Iowan(?).

Many small remnants of a glaciofluvial fill graded to the Iowan(?) fill occur high on the sides of the valleys of Willow Creek and Dry Run, tributaries of the Bad River (Pl. 1). This stratified drift is believed to have been derived from the Iowan glacier as it melted back to the east from its farthest westward extent.

The most continuous glacial sand-and-gravel deposit in the Missouri trench in the Pierre area, here informally referred to as the Oahe valley fill, is preserved in large terrace-topped remnants that range in altitude from 1530 feet to 1620 feet. This wide variation is caused partly by an elevation difference between different parts of a single terrace remnant due to differential compaction of the deposit, partly by a loess cover locally more than 30 feet thick, and partly by the stream gradient through a distance of more than 30 miles. In most places where this fill has been observed, it consists predominantly of sand and pebble gravel with numerous cobbles and some very large boulders. The presence of boulders as much as 8 feet in largest dimension seems to require that much of the fill accumulated when the ice margin was very close to, if not actually in, the Missouri River trench. Some of this deposit may actually be of ice-contact origin, although till inclusions and collapse features characteristic of such deposits have not been noted.

The Oahe valley fill has been traced north of the Oahe quadrangle to Spring Creek, a tributary that heads in the Pierre and Body Ranch end moraines. In the valley of Spring Creek, a body of outwash sand and gravel that appears to have been derived from the Tazewell(?) glacier during the building of the moraines is graded to the upper surface of the Oahe valley fill, which is thus the same age as the moraines and is therefore mapped as Tazewell(?). This general relationship was suggested by Rothrock (1944, p. 6) when he correlated an "ice-front" north of Snake Butte with the outwash in the valley of Spring Creek.

A group of small outwash deposits lies in the valley of Dry Run near the east edge of the Pierre quadrangle, and in the Canning quadrangle. Although the downstream equivalent of these deposits is lacking in the south-trending segment of Dry Run, the altitude of the remnants indicates that they are a part of a fill that was graded to the Oahe valley fill in the Missouri trench. This relationship is consistent with the view that the outwash was derived from the Tazewell(?) glacier, because Dry Run heads in the Tazewell(?) Body Ranch end moraine.

Tazewell outwash is not present west of the Missouri River. Several small deposits of stream gravel in the Bad River trench at altitudes of 1615 to 1640 feet possibly represent a former valley fill in the Bad River valley graded to the Oahe valley fill in the Missouri.

Drift borders.—The term drift border here refers to a line tangent to the outermost glacial drift deposited directly by ice. It does not include bodies of outwash which might extend far beyond the original outer margin of the ice sheet.

The end moraines of the Pierre area do not represent the maximum westward extent of glaciation, for Tazewell(?) ground moraine extends west to the Missouri River trench. West of the trench, scattered erratics and outwash deposits in tributaries of the Bad River prove that at least one ice sheet crossed the Missouri and extended some distance beyond it. A line connecting the outermost known glacial boulders is believed to represent the approximate westward and southward limits of glaciation (Fig. 5).

No ice-marginal drainage channels at or beyond the glacial drift border are known in this part of the State. Flint (1949, p. 69) suggested that meltwater from the ice sheet as well as discharge contributed by the major east-flowing rivers was ponded by the ice to form large lakes in the capacious valleys of these large tributaries of the Missouri River. Before these lakes overflowed and cut new channels near the ice front, the glacier retreated, uncovering the Missouri River trench.

Flint (1949) presented evidence that the glacial drift border west of the Missouri River is Wisconsin and probably is Iowan. There is no evidence that the Tazewell ice crossed the trench in the Pierre area, although the coarse nature of the Tazewell(?) outwash indicates that the glacier reached the trench at least locally.

Iowan-Tazewell Interval

The only event of apparent significance between the Iowan and Tazewell glaciations appears to have been downcutting in the Missouri River trench. The top of the Iowan(?) outwash fill surface, except for ice-contact stratified drift adjacent to Peoria Bottom, lies at an altitude of about 1700 feet. The base of the Tazewell(?) Oahe valley fill is about 150 feet below this altitude (Fig. 6). Between the deposition of these two fills, there evidently was 150 feet of degradation within the trench. The surface altitude of the lowland cut during the Sangamon age north of Snake Butte indicates that the Missouri trench had been lowered to at least 1500 feet before the Iowan sub-age. It seems likely that the Iowan(?) fill surface at 1700 feet in the trench was underlain by at least 200 feet of sand and gravel. Inasmuch as the dissection that took place before the Tazewell sub-age occurred primarily in unconsolidated fill, the difference in altitude between the Iowan and Tazewell surfaces does not carry much conviction as a record of an appreciable time interval. If both the Tazewell and Iowan substages are represented by glacial deposits in the Pierre area, there is no evidence of an appreciable time interval between them. Flint (manuscript in preparation) gives evidence that the Iowan-Tazewell break represents only an insignificant time interval in South Dakota.

Tazewell-Cary Interval

Field investigations in other areas in South Dakota and in adjacent states indicate that the time break between the Tazewell and Cary sub-ages was the most significant of the three Wisconsin interstadials. In the absence of more certain dating of specific glacial deposits in the Pierre area, little corroborative evidence has been found.

A qualitative examination of the surfaces of the Tazewell(?) end moraines and of the Cary end moraines in Hughes County reveals a morphologic difference that is very likely a

EXPLANATION
Outermost glacier limit (Iowan ?)
Probable Cary ice margin
Probable Tazewell(?) ice margin
Mankato ice margin
HYDE
SULLY
HUGHES
LYMAN
STANLEY
JONES
ARMSTRONG
PIERRE
MISSOURI
RIVER
FORT PIERRE
Bruns
Canning
Creek
Creek
Creek

function of the length of time these two drift sheets have been exposed to mass-wasting and stream erosion. The Tazewell(?) end moraines in the Pierre and Canning quadrangles have a fairly well integrated drainage pattern and there are few closed depressions. In contrast the surface of the Cary drift in eastern Hughes County is poorly drained and contains large numbers of closed depressions. Flint (manuscript in preparation) stated that these morphologies are characteristic of the surfaces of the Cary and Tazewell drift sheets in other parts of South Dakota. He also mentioned that the topographic expressions of the Tazewell and Iowan drift sheets are closely similar and that the topographic differences between the Cary and the Mankato drift sheets are slight. Accordingly, the evidence from the Pierre area is consistent with the thesis that the Tazewell–Cary interval was marked by conspicuous changes.

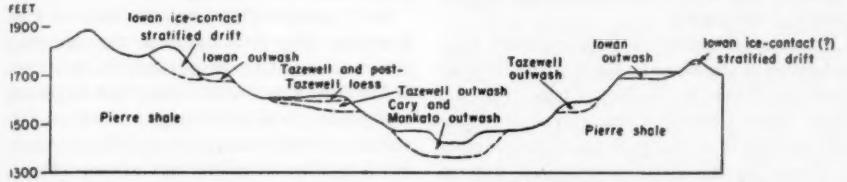


FIGURE 6.—IDEALIZED CROSS SECTION OF THE MISSOURI RIVER TRENCH IN THE PIERRE AREA
Showing inferred sequence and vertical distribution of stratified drift of the Wisconsin substages.

It is suggested subsequently that Cary(?) stratified drift underlies the surface of the modern Missouri River floodplain, thus raising the question of the time of cutting of the valley bedrock floor. At the time the Oahe valley fill was deposited, the Missouri River trench was nearly 2½ miles wide. At the site of the Oahe Dam, the maximum width of the bedrock-floored trench is about a mile. It seems likely that the Oahe valley fill was dissected and the bedrock floor of the trench was lowered from an unknown position to its present altitude during the Tazewell–Cary interval. This event represents downcutting of nearly 250 feet, about 150 feet of which probably was in Pierre shale. This lowering of the trench floor ascribed to the Tazewell–Cary interval records an event of considerable importance.

Cary(?) Glacial Drift

Till and moraines.—Deposits of the Cary substage in the area mapped in detail have been tentatively identified in the Canning quadrangle, although no single exposure of till can be definitely dated as Cary rather than Iowan or Tazewell. The surficial till in the northeast corner of the quadrangle is essentially identical in gross lithology with the Tazewell(?) till at the test pit and airport exposures. Nevertheless, several lines of evidence indicate that a post-Tazewell glacier reached the northeast corner of the Canning quadrangle.

Unoxidized till that grades upward into oxidized drift is exposed at several localities in and adjacent to the valley of Medicine Creek upstream from Canning. Whereas in these exposures the zone of oxidation generally is less than 20 feet thick, the oxidation zone in Tazewell(?) till to the west commonly is 40 feet thick where not overlain by later drift. North of Canning, ground moraine that retains constructional topography extends from upland areas down almost to the floodplain of Medicine Creek. Elsewhere in the Canning and Pierre quadrangles, till identified as Tazewell(?) in a corresponding topographic position has lost all semblance of constructional topography.

As the northeast corner of the Canning quadrangle lies nearly 15 miles southwest of the Mankato drift border mapped by Flint, the post-Tazewell(?) drift is inferred to belong to the Cary substage.

Stratified drift.—The valley of Medicine Creek contains remnants of a coarse Cary(?) outwash fill. Near Canning an extremely coarse deposit of outwash, consisting chiefly of cobbles and boulders, lies about 70 feet topographically

FIGURE 5.—WISCONSIN ICE MARGINS IN CENTRAL SOUTH DAKOTA
In part after Flint (manuscript in preparation).

below the upper surface of the Tazewell(?) fill terrace at the same locality. This vertical relationship suggests that the lower deposit most likely postdates the Tazewell(?) stratified drift and is Cary. The coarse grain size of the Cary(?) outwash seems to demand that it was deposited very near, if not immediately at, the margin of the glacier.

The nearest source of Cary outwash sand and gravel in the Missouri River upstream from Medicine Creek is Okobojo Creek (Fig. 5). From where Okobojo Creek leaves the Cary drift border, it is nearly 50 miles downstream to Pierre. In this distance, sediments derived from the Cary glacier should be well size-sorted with respect to increasing distance from the source. Therefore Cary stratified drift in the Missouri trench at Pierre, if present, should be fine-grained, with no cobbles and boulders.

Subsurface exploration at the site of the Oahe Dam and excavations for piers of the highway bridge across the Missouri River at Pierre indicate that the lowest point on the bedrock floor of the trench is at an altitude of about 1360 feet. The altitude of the floodplain in the same locality ranges from 1420 to 1440 feet. Thus a valley fill from 60 to 80 feet thick underlies the floodplain. Drill logs of this fill, provided by the Corps of Engineers, U. S. Army, indicate that the sediment is predominantly fine-grained, with approximately 90 per cent of the material sand size or smaller. Included in this fine-grained material are lenses of pebble gravel. Whereas the Missouri River at flood stage can probably transport all sizes of the alluvium that underlies the floodplain, at least the basal parts of this fill probably have remained undisturbed since original deposition. This is inferred from the discovery of the skull of a bison near the bottom of an excavation for a pier of the highway bridge. The skull shows no wear or breakage as would be expected had it been moved since the original deposition of the outwash.

Exposures of Cary outwash above the altitude of the Missouri River floodplain have not been identified with certainty. A deposit tentatively referred to the Cary substage lies adjacent to the Missouri River in the northwest part of the city of Pierre. This deposit consists predominantly of silt and fine- to medium sand, interbedded with lenses of pebble gravel. The upper surface of the outwash lies at an altitude of about 1440 feet. That this deposit is Tazewell(?) has not been disproved, but its relatively fine-grained nature is at least consistent with its tentative assignment to the Cary substage.

Drift border.—The westward limit of Cary glaciation (Fig. 5) is based on the occurrence of very coarse Cary(?) outwash in the trench of Medicine Creek at Canning, and on present topography of the drift surface. A large concentration of erratics along the east side of a north-trending ridge southeast of Canning (Fig. 5) suggests that this ridge baffled out much of the load carried in the basal parts of the ice sheets that crossed it. For a mile or more west of the ridge crest, drift is thin or scarce. The ridge probably was a barrier to the Cary(?) glacier; on the basis of this inference, the Cary(?) drift border is shown as trending along the east side of the ridge. North of Canning, the margin of the Cary(?) glacier probably extended west to an altitude of about 1700 feet and trended north, staying east of Medicine Knoll, to Blunt. The westward limit of the glacier north of Canning is only approximate and may be revised appreciably subsequent to more detailed work in that area.

Cary–Mankato Interval

In the area mapped, Mankato drift has been identified with reasonable certainty only in the valley of Medicine Creek. The vertical relationship of the Mankato(?) stratified drift to Cary(?) outwash at one locality suggests that there may have been 50 feet of downcutting between the two sub-ages. Dissection in outwash is not believed to be a reliable indicator of an appreciable time interval. From several lines of evidence, Flint (manuscript in preparation) concluded that the Cary–Mankato break was a very inconspicuous event in South Dakota. This same opinion was gained by Ruhe (cited in Flint) from an investigation of deposits of the four Wisconsin substages in northwestern Iowa.

Mankato Glacial Drift

The Mankato drift border trends northwest through the northeast corner of Hughes

County (Fig. 5), 15 miles or more east of the Canning quadrangle. Accordingly, Mankato till does not occur in the area mapped in detail. Flint (manuscript in preparation) noted that both the north and south branches of Medicine Creek contain Mankato stratified drift derived from a sublobe of the Mankato glacier in southeastern Sully County.

Half a mile upstream from the mouth of Medicine Creek, on the east wall of the valley, grading of a dirt road has exposed a deposit of silt and fine sand containing layers of medium sand and pebble gravel. A similar deposit lies at the about the same altitude on the west wall directly opposite. These deposits are about 25 miles downstream from the Mankato drift border. They lie nearly 50 feet topographically below Cary(?) coarse sand and gravel. On the basis of these facts, the lower, fine-grained stratified drift is believed to be of Mankato age. In most places in the valley of Medicine Creek, Mankato outwash has been covered by modern floodplain deposits and by products of mass-wasting processes.

Although Mankato stratified drift probably is present in the material underlying the Missouri River floodplain, it has not been differentiated from Cary(?) outwash sand and gravel.

Wisconsin Loess Deposits

Loess is widely distributed in the region surrounding Pierre. East of the Missouri River, loess as much as 30 feet thick mantles glacial sand and gravel, till, Pierre shale, and Recent alluvium. On upland areas west of the Missouri, the loess generally is 1–2 feet thick and is patchy.

As glacial deposits of the Wisconsin stage only have been recognized in the Pierre area, it is likely that no pre-Wisconsin loess is present. All the loess in the area probably is Tazewell and post-Tazewell. The terminology of loess sheets in this report follows the usage of Flint (manuscript in preparation):

Loess overlying Tazewell(?) drift and underlying a well-developed soil profile	Tazewell(?)
Loess overlying Tazewell(?) drift and extending to surface	Tazewell(?) and post-Tazewell
Loess overlying Cary drift or a well developed soil profile and extending to surface	Cary and post-Cary

Distinctive features to differentiate Tazewell(?) and post-Tazewell loesses were not found, although considerable time was consumed in drilling auger holes and inspecting vertical exposures of the loess. It is possible though not probable that a distinction can be made by laboratory analysis of the loesses. The orientation of the Missouri River, from which most of the loess apparently was derived, provides a source from which loess could be transported to the upland areas from several different directions.

There is a more or less consistent relationship between loess thickness and distance from the principal source area. It was noted previously that loess immediately adjacent to the Missouri River trench north of Snake Butte is as much as 20 feet thick. To the east, the average thickness diminishes to about 12 feet at 1½ miles, to about 8 feet at 3 miles, and to 3 feet at 5 miles. The end moraines present another factor influencing loess thickness; on them, slopes are relatively steep and much of the loess originally deposited on the tops and flanks of knolls has since been removed; in the swales, loess accumulation has been favored both by original deposition and by addition of material washed in from adjacent slopes. There is a general decrease in average grain size eastward from the Missouri trench. It appears evident from the thickness and grain-size distribution that most loess has been derived from the Missouri River trench.

In many places in the Pierre area, one or more buried immature soil profiles in loess have been observed. Most of the profiles consist of the following layers or zones:

Upper: oxidized, partly leached, humified material

Lower: oxidized, unleached material; may be enriched in secondary calcium carbonate

Parent material: oxidized, unleached loess

In all outcrops and in nearly all auger holes, the loess is oxidized throughout. In several places, however, loess in a buried soil profile

is more oxidized than the loess between the buried profile and the present surface.

The time importance of the soil profiles of varying degrees of development is not known except in qualitative terms. The more strongly developed soil profiles are thought to record the relatively longer time intervals between periods of loess accumulation. It is unlikely that a "profile" that consists only of a buried humified zone has much time importance. There is some evidence that, under optimum conditions, such a zone could develop in a few decades.

In many of the loess sections overlying Tazewell(?) glacial drift in the Pierre area, there is one more or less well-developed, buried soil profile. This consists in most places of an incompletely leached zone, the only additional feature to set that layer apart from the loess above and below being a humified zone 4 to 18 inches thick. Less commonly the humified layer is underlain by a zone of secondary calcium carbonate. This is a more mature soil profile than those developed in loess overlying the Cary or Mankato drift sheets elsewhere in South Dakota. During the past several years, evidence has been steadily accumulating that the most important time break between successive glaciations during the Wisconsin was the Tazewell–Cary interval (Warren, 1952; Flint, manuscript in preparation; Ruhe, 1950). Therefore it is likely that the more mature profile in loess overlying the Tazewell drift in the Pierre area was developed in the Tazewell–Cary interval.

During the Tazewell sub-age, in areas within or close to the Missouri, as much as 11 feet of loess accumulated, whereas several miles from the trench only 2 to 3 feet were deposited. These figures are based on thickness of loess underlying and including buried soil profiles. Cary(?) and post-Cary loess, where separated from Tazewell(?) loess by a buried soil profile, is up to 10 feet thick within the Missouri River trench.

Summary of Pleistocene History

In the early part of the Pleistocene epoch, the surface of the Pierre area probably stood at altitudes ranging from 2200 to 2400 feet. Drainage was accomplished by east-flowing streams; the valley floor of one of these streams coincided with the crest of Standing Buttes. It is possible that this river was the ancestor of the modern Cheyenne River. The location of the ancestral Bad River before Yarmouth time is not known.

Physical evidence of the Nebraskan and Kansan glaciers is not found in the Pierre area. Presumably from the Nebraskan age to the late Kansan, the area was being slowly lowered by stream erosion and mass-wasting. Although downcutting very likely alternated with aggradation in the major east-flowing rivers, a progressive lowering of stream profiles took place.

By late Kansan or early Yarmouth time, the ancestral Bad River was flowing across the area in a channel about 6 miles north of the present Bad. During the Yarmouth, the Bad River progressively cut down and planed south, leaving veneers of alluvium on abandoned straths.

At the beginning of the Illinoian age the ancient Bad River was flowing at an altitude of about 1750 feet. Upland surfaces were smoothly graded to the river. Topographic inversion, caused by protective caps of stream alluvium left along former and higher channels of the Bad River, had taken place. As the Illinoian glacier advanced into central South Dakota, meltwater was ponded ahead of it in the major east-trending valleys. The meltwater, added to the normal discharge of the east-flowing streams, spilled across interfluves and cut a continuous channel when the margin of the Illinoian glacier had reached the approximate position of the present Missouri.

This cutting of the Missouri River trench rejuvenated the pre-existing streams of the area. Stream erosion and mass-wasting during the Sangamon age enlarged the trench to its present width and nearly to its present depth. So far as known, all glacial deposits of the Illinoian stage were removed during this period of erosion.

In the Wisconsin age, the Iowan glacier crossed the Missouri River trench and extended over nearly the entire Oahe quadrangle. Before the meltwater could again overflow and cut new spillways through interfluves, the ice

margin retreated to a position east of the Missouri trench. In the course of this eastward retreat, meltwater was discharged down valleys tributary to the Bad and built an outwash fill in these valleys and in the trench of the Bad River. This fill was graded to an outwash fill in the Missouri River trench. Deposits of ice-contact stratified drift originated adjacent to large masses of ice that lay stagnant in the Missouri trench. Meltwater from this ice mass and from the glacier margin east of the trench formed an outwash fill in the Missouri, the upper surface of which was at an altitude of about 1700 feet. The Iowan drift filled in and smoothed out the pre-existing rough topography formed by dissection in the Sangamon age.

Although evidence of an appreciable time interval between the Iowan and Tazewell sub-ages is lacking in the area, at least 150 feet of downcutting occurred in the Iowan outwash fill in the Missouri trench during this interval.

The advance of a glacier in the Tazewell sub-age further filled in and smoothed out pre-existing topographic irregularities. Although the Tazewell ice evidently pushed into the Missouri trench at most places in the three quadrangles, there is no evidence that it moved onto the upland south and west of the river. Soon after it reached its outermost limit, the Tazewell glacier melted back several miles. A condition of balance was then reached between wastage and flowage and the Pierre end moraine was built. After another brief period of retreat the ice margin again halted and the Body Ranch end moraine was formed. Meanwhile, meltwater from the Tazewell ice flowing into the Missouri by way of Spring Creek and Dry Run as well as from masses of ice in the trench itself built a thick and continuous outwash fill in the Missouri. When the Tazewell glacier melted back to Medicine Creek, the valley of that stream also was partly filled by outwash.

With the disappearance of the Tazewell glacier from the Pierre area, dissection started in the outwash fills. Very likely the present bedrock floor of the trench was cut in the Tazewell–Cary interval. Winds deflated silt from the Missouri floodplain and deposited the material on terraces in the trench and on adjacent upland surfaces. Before a glacial climate had again set in, oxidation, leaching, and humification produced an immature soil profile in the Tazewell loess.

The Cary glacier failed to extend as far west as did the Tazewell glacier, and the outermost edge of the ice sheet reached only the northeast corner of the Canning quadrangle. Stratified drift derived from this ice margin was deposited in the trench of Medicine Creek, where it now occurs as terrace remnants. It is likely that the Missouri River aggraded to an elevation at least 20 feet above the modern floodplain in response to the contribution of Cary stratified drift.

The maximum of the Cary substage was accompanied and followed by deflation of silt from Cary drift and the deposition of this material as loess. Before the maximum of the following Mankato sub-age, dissection of the Cary outwash fill had taken place.

The margin of the Mankato glacier reached northeast Hughes County, and the only Mankato glacial deposits in the quadrangles mapped are bodies of fine-grained outwash. So far as known, loess deposition has been more or less continuous from the Mankato maximum to the present.

The surface of the Pierre area now is being lowered by stream erosion, deflation, and mass-wasting processes. Because of the relatively impermeable character of most of the materials at the surface, and because of low precipitation, chemical weathering is taking place at a slow rate. Modern soil profiles in loess are less well developed than the profile developed during the Tazewell–Cary interval. This suggests that either a longer time existed between the Tazewell and Cary sub-ages than from the Mankato maximum to the present, or that the Tazewell–Cary break was characterized by a warmer and moister climate than the present one, or both.

References Cited

Crandell, D. R. (1950) *Revision of Pierre shale in central South Dakota*, Am. Assoc. Petr. Geol., Bull. vol. 34, p. 2337–2346.

——(1951) *Preliminary report on the geology of the Pierre quadrangle, South Dakota*, U. S. Geol. Survey, open file report, 28 pages.

Flint, R. F. (1949) *Pleistocene drainage diversions in South Dakota*, Geograf. Ann., p. 56–74.

——(manuscript in preparation) *Pleistocene geology*

of eastern South Dakota, U. S. Geol. Survey, Prof. Paper.

Rothrock, E. P. (1944) *Sand and gravel deposits in the Missouri Valley between Little Bend and White River, South Dakota*, Geol. Survey, Rept. Invest. 47, 118 pages.

Ruhe, R. V. (1950) *Graphic analysis of drift topographies*, Am. Jour. Sci., vol. 248, p. 435–443.

Todd, J. E. (1914) *The Pleistocene history of the Missouri River*, Science, new ser. vol. 39, p. 263–274.

Warren, C. R. (1952) *Probable Illinoian age of part of the Missouri River*, Geol. Soc. Am., Bull., vol. 63, p. 1143–1156.

U. S. Geological Survey, Geologic Division, Denver Federal Center, Denver, Colo.

Manuscript Received by the Secretary of the Society, June 16, 1952.

Publication Authorized by the Director, U. S. Geological Survey.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 64, PP. 599-600 MAY 1953

Short Notes

CAUSES OF BASALTIC EXPLOSIONS

BY HAROLD T. STEARNS

Extensive studies of basalts in Australia, New Zealand, Java, the Philippines, Japan, Oceania, the United States, Mexico, and Italy have convinced the writer that plateau and oceanic basaltic magmas are not explosive and that such explosions as occur are due to extraneous causes.

It is essential to distinguish plateau and oceanic basalts from basalts of unusual chemical composition and from those approaching andesites or trachytes. The latter are more viscous than the oceanic and plateau basalts and do not occur as pahoehoe flows.

Explosion, as used herein, refers to the paroxysmal eruption of accessory and accidental ejecta derived from foundation rocks with, or without, accompanying magmatic ejecta. It does not include the usual firefountains of vitric material which result solely from the normal expansion of magma upon reaching the surface.

Explosions may be classified as magmatic, phreatic, and phreatomagmatic. Phreatic explosions are caused by the addition of surface or ground water to hot rock, and no new magmatic material is present. The explosions at Kilauea in 1924, due to the infiltration of ground water, are typical of the phreatic type. No sharp line of distinction exists between the initial outbreaks of a new geyser and phreatic explosions at volcanoes, except in magnitude.

Phreatomagmatic explosions are those which occur when hot magma comes into contact with surface or ground water. Two types occur, one where water enters a vent and the other where lava flows into water. The products of these two types are readily distinguished in the field by the lack of appreciable quantities of ejecta composed of basement rock in deposits formed when lava runs into water and explodes.

Island volcanoes, during their early history above sea level, are characterized by phreatomagmatic explosions, but these explosions gradually decrease in number as the volcanoes build ever higher cones above the sea. In contrast, explosion deposits are extremely rare in the great lava plateaus of the world where they probably make up less than .1 per cent of the total volume of volcanic rock.

Geologists unfamiliar with the high permeability of basalts commonly do not suspect the large quantities of ground water available which start and sustain explosions. Many explosive vents occur on dry land where former high stands of the sea either provided the necessary water or raised the water table into permeable formations through which the magma erupted.

Near Roberts, Idaho, a fissure eruption crossed the coarse, highly permeable gravel fan of the Snake River and produced a chain of cones. Vents on the fan were violently explosive, and the cones contain 50 per cent or more gravel, but those just off the fan are composed chiefly of cinders.

A study of other basaltic explosion deposits in relation to the availability of surface or ground water is needed. Most writers describe the deposits but offer no explanation of the explosion. It is hoped that this note will encourage other geologists to record examples of explosions which may confirm or deny the hypothesis herein offered—namely, that plateau and oceanic basaltic magmas are nonexplosive and that extraneous causes, chiefly water, cause paroxysms.

HOPE, IDAHO

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 64, PP. 601-602 MAY 1953

UNIFORMITY OF THE EARTH'S MANTLE

BY FRANCIS BIRCH

The Earth's mantle is here taken to mean the thick layer extending from the base of the crust (Mohorovičić discontinuity) to the boundary of the central core (Oldham-Gutenberg discontinuity). The structure and chemical composition of this region are of fundamental importance for geological theory. A recently proposed method of approaching these problems (Birch, 1952) leads to the tentative conclusions that the region between depths of 900 and 2900 km may be effectively homogeneous, but that between about 200 and 900 km there are gradual changes of composition, or of phase, or both. These conclusions raise difficulties for the theories of convection currents in the mantle, and it may be useful to examine one consequence of the assumption that, on the contrary, the whole mantle is uniform in composition and phase, the change of density being determined by pressure and temperature alone.

Let us first consider adiabatic compression. The ratio of adiabatic incompressibility to density may be found from the seismic velocities, as was shown 30 years ago by Williamson and Adams (1923), and it is thus possible to find, by numerical integration, the change of density with radius on the assumption that it is determined by adiabatic compression. The relation is

$$d\rho/\rho = -g d\tau/\phi, \tag{1}$$

where $g = GM_r/r^2$, the acceleration of gravity at radius r, and $\phi = V_P^2 - (4/3)V_S^2$, to be found from tables of the velocities, V_P and V_S, as functions of radius. The procedure is to choose an initial density, ρ_0, for the top of the mantle and to integrate inward to find the density at all levels. An acceptable solution for the whole Earth must give the known mean density and moment of inertia. Without obtaining a complete solution, however, we may follow a method employed by Bullen (1936; 1940) to exclude certain choices of initial density by examining what is left of the total mass and moment of inertia of the Earth after the corresponding quantities for the mantle have been subtracted. After a small correction for the crust, the remaining mass and moment of inertia are those of the core, say M_c and I_c. The density of the core increases with depth because of its own compression, and it seems reasonably certain that the ratio, $z = I_c/M_cR_c^2$, where R_c is the radius of the core, must be less than 0.4, the value of this ratio for a sphere of uniform density. Thus any distribution of density for the mantle which leads to a value of z of more than 0.4 must be rejected.

Bullen has solved this problem for a particular value, $\rho_0 = 3.32$, suggested by the density of the Moon and the density of ultrabasic rocks such as dunite. On the assumption of adiabatic compression of a homogeneous layer, according to (1), Bullen found that this initial density led to the unacceptable value, $z = 0.57$, for the core. An acceptable solution was then found by introducing an arbitrary increase of density, first at a single level (1936), then distributed through a region (1940). This solution implies nonuniformity of some kind: the density is no longer determined by adiabatic compression alone.

This important result of Bullen's calculation has not received the attention it deserves, possibly because it has been thought that a minor change of premise would alter the conclusion. It is thus of interest to have comparable results for a wide range of initial density. The computation is much simplified, while retaining adequate precision for the present problem, by neglect of the variation of g with depth; this variation is known to be small for all the modern density distributions. With g constant throughout the mantle, we have

$$\log \rho/\rho_0 = g\int_r^R dr/\phi \tag{2}$$

where R is the outer radius of the mantle. The computation has been carried through for two different values of g: 1000 gals and 980 gals. The velocities were taken from Jeffreys (1939); the allowances for the mass and moment of inertia of the crust, taken as 33 km thick, are those of Bullen (1940). The results are shown in the following table.

	$\rho_0 = 3.2$	3.3	3.4	3.5	3.6	3.7	3.8	3.9
$g = 980$,	$z = 0.62$	.58	.54	.49	.44	.38	.32	.25
$g = 1000$,	$z = 0.61$	.57	.53	.48	.43	.37	.30	.23

From these figures, it follows that, if the change of density in the mantle is determined by adiabatic compression alone, then the initial density at the top of the mantle cannot be much less than 3.7. Allowance for a rate of rise of *temperature* with depth greater than the adiabatic rate reduces the rate of increase of *density* with depth below that given by adiabatic compression, so that an even higher initial density is required. In the limiting case of uniform density (Wiechert's hypothesis), the required density is about 4.2. Bullen's adjusted density distributions correspond to values of z between about 0.37 and 0.39.

We have little direct knowledge of the density below the crust, but there are reasons for thinking that 3.7, at a depth of 33 km, is improbably high. An olivine of this density would be iron-rich, with velocities too low to meet seismic requirements. An eclogite might meet these requirements (Adams and Gibson, 1929), but a complete mantle of eclogite seems implausible from the standpoint of chemical abundances. With any initial density less than about 3.7, it is necessary to introduce an increase of density, somewhere in the mantle, in addition to the effect of compression. This can be done in different ways, and attributed to different causes, but in view of earlier discussions (Birch, 1939; 1952; Bullen, 1940) it appears most reasonable to associate the density increase with the anomalous rise of velocities in the region between depths of 200 and 900 km, as Bullen has done. The amount of increase required depends upon the choice of initial density, for which a somewhat higher figure than 3.32 may eventually be preferred.

References Cited

Adams, L. H., and Gibson, R. E. (1929) *The elastic properties of certain basic rocks and of their constituent minerals*, Nat. Acad. Sci., Pr., vol. 15, p. 713–724.

Birch, Francis (1939) *The variation of seismic velocities within a simplified earth model, in accordance with the theory of finite strain*, Seismol. Soc. Am., Bull., vol. 29, p. 463–479.

—— (1952) *Elasticity and constitution of the Earth's interior*, Jour. Geophys. Res., vol. 57, p. 227–286.

Bullen, K. E. (1936) *The variation of density and the ellipticities of strata of equal density within the Earth*, Mon. Not. Royal Astr. Soc., Geophys. Supp. 3, p. 395–400.

—— (1940) *The problem of the Earth's density variation*, Seismol. Soc. Am., Bull., vol. 30, p. 235–250.

Jeffreys, Harold (1939) *The times of* P, S, *and* SKS, *and the velocities of* P *and* S, Mon. Not. Royal Astr. Soc., Geophys. Supp. 4, p. 498–533.

Williamson, E. D., and Adams, L. H. (1923) *Density distribution in the Earth*, Washington Acad. Sci., Jour., vol. 13, p. 413–428.

Dunbar Laboratory, Harvard University, Cambridge, Mass.